A Tectono-Geomorphic Study of the Alpine Fault, New Zealand

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ABSTRACT

The Alpine Fault is a ~900 km-long, active Australian-Pacific plate boundary structure, which accommodates up to 70–90% of total plate boundary motion across the South Island of New Zealand. Despite abundant evidence that large to great (≈M 8) magnitude earthquakes have occurred frequently and regularly on the fault in the past, it has not ruptured historically and is thought to pose one of the greatest seismic hazards to the country of New Zealand at present.

This study adopts a multi-disciplinary field-based approach to examine fault zone structure and mechanics, spatio-temporal variations in fault behavior, and geomorphic evidence of key coseismic hazards on the central and southern Alpine Fault. The first three complete sections through the fault core of the southern Alpine Fault show that modern slip is localized to a single 1 to 12 m-thick fault core composed of impermeable \( (k = 10^{-20} \text{ to } 10^{-22} \text{ m}^2) \), frictionally weak \( (\mu_s = 0.12–0.37) \), velocity-strengthening, illite-chlorite and saponite-chlorite-lizardite fault gouges. The frictionally-weakest fault gouge occurs in the widest fault core and is spatially associated with a newly-identified serpentinite-bearing tectonic mélangé.

In contrast to the relatively straight and localized dextral-normal-motion fault traces of the southern Alpine Fault, the central Alpine Fault is characterized by non-optimally-oriented oblique dextral-reverse motion, which causes the fault zone to partition in the upper ~1–2 km. Utilizing airborne light detection and ranging (LiDAR) data, the surface expression of a portion of the central Alpine Fault was mapped in unprecedented detail to confirm previous mapping that shows the fault is composed of serially-partitioned (i.e., sequenced) oblique-thrust and strike-slip faults at 1–10 km-length scales, and introduce for the first time the widespread occurrence of ~300 m-wide parallel-partitioned positive flower structures. A fault kinematic analysis predicts the fault trace orientations observed and supports the concept that the partitioning behavior is scale dependent, with different mechanisms (i.e., crustal-scale discontinuities, thermal weakening, fluvial incision, sediment interaction) exerting control at different scales (~10–10 m). A slip stability analysis suggests that the newly-formed shallowly-rooted faults are kinematically stable, and thus the existing ~300 m-wide zone of fault traces defines a surface rupture hazard zone where future ruptures are expected to occur.

Deep-seated, long runout, catastrophic rock avalanches currently represent an underappreciated hazard of Alpine Fault earthquakes. The previously undescribed ~0.75 km³ c. 660 AD Cascade rock avalanche has an unambiguous structural relationship to pre-existing deep-seated bedrock failures. In comparison with other documented rock avalanches in the Southern Alps and Fiordland, it provides clues about precursory conditions for large catastrophic failures and suggests a mass above Franz Josef (town) poses a considerable risk.

A remarkable ~8 km dextral offset of major valleys and glacial deposits is recorded along ~100 km of the southern Alpine Fault. Tight age constraints allow correlation of this event to the Waiauunga Glaciation (Marine Isotope Stage 8; c. 270 ka) and indicate a dextral Alpine Fault slip-rate of 29.6 (-2.1/+2.3) mm/yr. Ages of marine sediments uplifted to ~600 m elevation yield fault-proximal Australian plate uplift rates of ~2.2–2.5 mm/yr. A reassessment of the slip-rate and uplift rate catalog for the southern Alpine Fault suggests relatively constant rates over the last > 300 kyrs, and potentially > 3.5 Myrs.

Together, the results of this study frame a view of the southern half of the Alpine Fault as a highly-localized, long-lived, very weak locus of plate boundary motion that has had relatively constant spatio-temporal displacement rates in the latter part of its history, ruptures in hazardous large magnitude earthquakes with strong peak ground accelerations, and exerts a first-order control on landscape evolution of the South Island.
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To all of you and any forgotten, my deep and sincere gratitude.
“The design of a book is the pattern of a reality controlled and shaped by the mind of the writer. This is completely understood about poetry or fiction, but it is too seldom realized about books of fact. And yet the impulse which drives a man to poetry will send another man into the tide pools and force him to try to report what he finds there.”

– John Steinbeck (1902–1968), Writer

“Nothing is less real than realism. Details are confusing. It is only by the selection, by elimination, by emphasis that we get at the real meaning of things.”

– Georgia O’Keeffe (1887–1986), Painter

“Civilization exists by geological consent, subject to change without notice.”

– Will Durant (1885–1981), Philosopher/Historian

“In that glorious future to which Westcoasters are always looking forward to, when the ‘Something is bound to turn up’ does turn up, when mines will be opened of fabulous richness, and when industries yet undreamed of will bring peace and prosperity, Churches, Gaols & the Tax Collector to South Westland, there may be a Township formed at the mouth of the Cascade – Sandfly would be a good name.”


“There is nothing as sobering as an outcrop.”

– Francis Pettijohn (1904–1999), Geologist

Dedicated to you, the reader. The oft unsung hero in this. May you find knowledge, inspiration and provocation in these pages.
Oblique aerial photograph looking northeast along the Alpine Fault (vertical at center) from above the Jerry River. Low Creek lateral moraines (left foreground) are spaced ~1 km apart. Photo taken by L. Homer (CN14548/34). Photo courtesy of GNS Science.

“Ngā kīnga a Papatūānuku atu, ki a ngā whakaoko”
Annotated oblique aerial photograph looking northeast along the Alpine Fault (vertical at center) from above the Jerry River. Annotations highlight aspects of research introduced and examined in this thesis. Low Creek lateral moraines (left foreground) are spaced ~1 km apart. Photo taken by L. Homer (CN14548/34). Photo courtesy of GNS Science.

“The Earth speaks to those who listen.” – Anon.
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Chapter 1  Introduction

Figure 1.0    Oblique aerial photograph looking southeast across the Cascade valley near Woodhen Creek from above Mt Delta. The Cascade River flows to the left at bottom and the Alpine Fault traverses horizontally at center. In contrast to the better-studied central section of the Alpine Fault which exhibits oblique-reverse offset, the southern Alpine Fault has pronounced characteristics of an almost-purely strike-slip fault. Notice here the presence of sag ponds, anticlinal ridges and dextrally offset creeks and alluvial fan surfaces. This portion of the Alpine Fault juxtaposes particularly diverse lithologies; here Greenland Group metasedimentary rocks (bottom) are juxtaposed against Dun Mountain-Maitai Group peridotite (top). Distance between the Alpine Fault and the Cascade River is ~800 m. Photo taken by L. Homer (CN5258/B). Photo courtesy of GNS Science.
1.1 Aims

This study aims to contribute to an understanding of the tectonophysics, geomorphology and geological hazards related to the southern and central portions of the Alpine Fault, the main Australian-Pacific plate boundary structure traversing the country of New Zealand (Figure 1.1). Along with Turkey’s Northern Anatolian Fault and California’s San Andreas Fault, New Zealand’s Alpine Fault is one of the world’s great active continental strike-slip faults. Unlike these other two structures, the Alpine Fault has not had a large magnitude surface-rupturing earthquake since European settlement so the effects and extent of such an event have to be estimated solely from naturally-preserved records in the landscape (e.g., fault scarps, disturbed trees, landslides, river aggradation events). A long (8000 yr; 24 earthquakes) paleoseismic record (albeit from a limited number of sites) of quasi-periodic surface-rupturing earthquakes on the fault, in conjunction with some along-strike control on the surface rupture lengths of the last few major earthquakes, indicate the Alpine Fault is late in its seismic cycle and capable of generating a ~Mw 8.0 earthquake [Sutherland et al., 2007a and references therein; Wells and Goff, 2007; Berryman et al., 2012a]. With the memory of the 2010–2011 Christchurch earthquakes still firmly imprinted on the entire country’s consciousness and the looming 30 % probability of a ~Mw 8 Alpine Fault earthquake in the next 50 years [e.g., Sutherland et al., 2007a; Berryman et al., 2012a], now is the time to learn as much as possible about the physics governing the Alpine Fault zone and the extent of geological hazards that will be associated with this future major earthquake.

The topics outlined in this thesis are explored using data, samples and ideas collected in the course of fieldwork into remote areas along the southern Alpine Fault, one of the structurally most complicated and lithologically most diverse regions in New Zealand. This thesis explores several key questions over the course of four main chapters:

What are the anatomy and physical/chemical/mechanical properties of the southern Alpine Fault and how do these differ from the better-studied central Alpine Fault? What effects do lithologies play on fault zone properties? How has the structure evolved with time? What can its nature tell us about earthquake behavior? (Chapter 2)

How wide and variable is the Alpine Fault surface rupture zone? Do ruptures occur repeatedly in the same place or are new fault scarps formed in each earthquake event? What can detailed surface morphology and outcrop data reveal about the three dimensional structure of the fault at shallow depths? What are the key factors controlling how plate
boundary slip is partitioned onto different near-surface structures at length scales from ~1000 km to ~1 m² (Chapter 3)

What is the rock avalanche hazard associated with major Alpine Fault earthquakes? How does this compare with hazard due to earthquakes on other structures or other rock avalanche triggers? Can deposits from past large rock avalanches be used to date earthquakes and estimate co-seismic ground shaking? What pre-existing conditions and bedrock structures can be used to identify sites of potential catastrophic failure? (Chapter 4)

How constant have uplift and strike-slip rates been on the southern Alpine Fault through time? How do these rates compare to those determined to the north and further south? What history of landscape evolution (e.g., glaciations, drainage captures, relative sea levels) is preserved, and how does this relate to dextral offset and vertical uplift on the southern Alpine Fault? (Chapter 5)

This thesis employs an integrated approach using diverse techniques to reveal and highlight new aspects of the tectonophysics, geomorphology and geologic hazard of the Alpine Fault. The study broadly draws on basic field techniques from structural geology, geomorphology, petrology and stratigraphy, but within it I have performed wide-ranging pilot studies testing and incorporating more detailed advanced methods such as microstructural determination of crystallographic preferred orientations (CPO) using electron backscatter diffraction (EBSD), energy-dispersive x-ray spectroscopy (EDS), x-ray diffraction (XRD), fault rock friction experiments, carbonate stable isotope analyses, \(^{40}\text{Ar}/^{39}\text{Ar}\) radiometric dating, fault stability analyses, examination of interferometric synthetic aperture radar (InSAR) data and airborne light detection and ranging (LiDAR) data, stream profile analyses using ArcGIS and Matlab programs, amino acid racemization (AAR) dating, and even nannopaleontology. In my opinion the impact of the integrated story told by these studies far outweighs a more detailed study of any of its parts. As this thesis demonstrates, even a 5 µm-wide nannofossil can have a staggeringly important tectonic story to tell when paired with careful stratigraphic and geomorphic observations. This versatility of applications is central to the goal of this thesis to achieve far-reaching, broader picture views of this most remarkable plate boundary structure, the Alpine Fault.

1.2 Timeliness of the Thesis

Several coeval major geological investigations provide timely comparisons to this study. Starting in 2004, the main phases of the San Andreas Fault Observatory at Depth (SAFOD)
drilled a scientific borehole through an actively creeping portion of the San Andreas Fault near Parkfield to a depth greater than 3 km. Results from this program presented in the last few years [e.g., Bradbury et al., 2011; Holdsworth et al., 2011; Lockner et al., 2011; Zoback et al., 2011 and references therein; Moore and Rymer, 2012; Schleicher et al., 2012] have led to a small renaissance in the understanding of the processes of rock deformation and earthquake physics operating in an active plate boundary fault at depth. A similar project on the central Alpine Fault, the Deep Fault Drilling Project (DFDP), has completed its first phase in 2011 and plans to drill a ~1.5 km deep scientific borehole through the fault in 2014 [Townend et al., 2009; Sutherland et al., 2012] with a longer term goal of drilling to 5 km depth. Results from these two projects can be compared to those in the present study to elucidate the structure of major continental fault zones and their spatial variation (Chapter 2). LiDAR interpretation made as part of the present study (Chapter 3) has helped borehole locations to be chosen in both phases of the DFDP project. The present study also benefits from recently cataloged paleoseismic studies [Berryman et al., 2012a; Howarth et al., 2012], which have greatly improved our understanding of the seismic behavior of the Alpine Fault and to which my results can be compared.

The field data gathered in this study were incorporated into the 1:250,000 scale geological map of the Haast region published by GNS Science [Rattenbury et al., 2010], which encompasses the northern half of the main study area of this thesis. The published map incorporates all my field observations to the date of publication. While a few minor changes to that map are indicated by the most recent field observations during this present study, and a few more significant changes are indicated as needed for the map of the southern half of the main field area (Wakatipu 1:250,000) [Turnbull, 2000], based on the overall accuracy and quality of both these maps, there is little need to produce a comprehensive geological map of the region from my study. Where applicable, key differences in mapped geology introduced by this thesis are highlighted in text or figures (and all field data are presented in a digital appendix).

The 2010 and 2011 Christchurch earthquakes were, and continue to be, potent reminders that New Zealand is positioned astride an active plate boundary subject to periodic large and damaging earthquakes. The earthquakes in Christchurch have had far-reaching psychological and economic effects on the entire country of New Zealand and have driven many regional councils and city planners to re-assess their risk and impact of future earthquakes. In particular, the West Coast Regional Council must continue their strategies to plan and
prepare for the likelihood and consequences of a major Alpine Fault earthquake (30% chance in the next 50 years) [Berryman et al., 2012a]. Mapping of active fault traces along the Alpine Fault has allowed definition of a buffer zone around the mapped surface ruptures in which buildings should not be built and from which existing buildings should be moved [Langridge and Ries, 2010]. LiDAR interpretation in the present study has helped refine the mapped surface rupture zone in the vicinity of Franz Josef township [Langridge and Beban, 2011; Barth et al., 2012; Chapter 3]. A re-assessment of Alpine Fault-triggered rock avalanche hazard (Chapter 4) also contributes to a growing body of work by concerned scientists aiming to mitigate the forthcoming impacts to humanity [see also Orchiston, 2012; Robinson and Davies, 2013].

1.3 Thesis Structure

Only a brief introduction to New Zealand geology is provided in the present chapter. More comprehensive separate introductions are given in the subsequent chapters to highlight aspects of the Alpine Fault most relevant to the particular topic. Following this introductory chapter (Chapter 1), the results of this PhD are divided into four main chapters. Chapter 2 introduces and defines the southern Alpine Fault and provides details of the physical and mechanical properties of the fault zone. These data provide a useful comparison to results from the Deep Fault Drilling Project (DFDP) on the central Alpine Fault and to other major strike-slip faults worldwide. Chapter 3 focuses on the central Alpine Fault, where interpretation of LiDAR data is used to understand how the way the plate boundary is partitioned at various scales relates to thermal weakening, topography, fluvio-glacial incision and sedimentation; additionally the data are used to define a surface rupture hazard zone for the fault. Chapter 4 focuses on the southern portion of the fault where the very large, catastrophic, Cascade rock avalanche is found in the Cascade valley. Study of this feature improves the Alpine Fault paleoseismic record and I argue that such rock avalanches may be an underappreciated hazard of a forthcoming major Alpine Fault earthquake. I also predict locations and conditions in which future catastrophic events may be likely to occur along the fault and elsewhere in the Southern Alps. Chapter 5 explores the remarkable history of greater than 300 kyrs of landscape evolution preserved on the Australian plate by dextral-normal motion on the southern Alpine Fault; in addition to providing constraints on mid-range Alpine Fault slip rates and uplift rates in the area, it also serves as an important record of glaciations not preserved elsewhere in New Zealand and allows estimates for the tempo of various landscape events. The implications and conclusions of the previous four chapters are
then brought together in a brief chapter synthesizing the study and outlining suggestions for future work (Chapter 6).

To maintain a clear and concise presentation of key topics, only essential data are incorporated into the text of the thesis chapters. Important, but slightly tangential data, figures and text discussions are included as auxiliary materials in the Supplements. These Supplements are arranged according to the corresponding chapter (e.g., the in-text citation for the Supplement to Chapter 2 is “Supplement 2”). The Appendix is used primarily for a catalog of sample names, supporting data tables, and paleontological reports. Two DVDs of digital data is also provided. In addition to materials within the bound thesis, the digital data includes all field notes (both scanned and typed georeferenced), raw data files, many field photographs, conference posters, and ArcGIS databases containing various geospatial datasets with the hope that it aids or inspires future research.

1.4 New Zealand Geology

1.4.1 Recent tectonics

New Zealand comprises the ~10% of the continent of Zealandia presently emergent from the Pacific Ocean; in total the area of continental crust (defined as where bathymetric depth is < 2000 m) is roughly half the size of Australia. Presently New Zealand is situated astride the Australian-Pacific plate boundary which locally experiences long-term relative plate motion rates at rates of ~35–40 mm/yr (Figure 1.1). The most obvious onshore manifestation of this plate boundary in New Zealand is the Alpine Fault, the trench-to-trench transform which links northeast-directed subduction of the Australian plate beneath the South Island (the Puysegur subduction zone) to west-directed subduction of the Pacific plate beneath the North Island (the Hikurangi subduction zone). Both subduction zones accommodate oblique motion on parallel-partitioned thrust and strike-slip structures. Much of the strike-slip component is accommodated on trench-parallel strike-slip fault systems (e.g., southernmost Alpine Fault, North Island Fault System) [e.g., Barnes et al., 2005; Nicol and Wallace, 2007 and references therein]. The Alpine Fault accommodates 60–90% of plate boundary motion on its southern and central portions [Norris and Cooper, 2001; Barnes, 2009], and the northernmost portion (commonly referred to as the Wairau Fault) has been largely abandoned in favor of < 7 Ma strike-slip faults of the Marlborough Fault System [Little and Jones, 1998] (Figure 1.1). Motion across the Marlborough Fault System has generally been migrating southward with time onto newer-formed and better-oriented faults that link with
Figure 1.1  New Zealand Tectonics. Key tectonic features of the Australian-Pacific plate boundary in New Zealand highlighting the context of the study areas. Northwest-illuminated hillshade derived from a digital elevation model (DEM) of Land Information New Zealand (LINZ) topographic data and National Institute of Water and Atmospheric Research (NIWA) 250 m bathymetry data. Australian-Pacific relative plate motion vectors (in mm/yr) from DeMets et al. [1994]. Subduction zone is abbreviated as S.Z.

the southward lengthening subduction margin [Little and Roberts, 1997; Little and Jones, 1998; Wallace et al., 2012]. The southernmost of the through-going Marlborough faults, the Hope Fault, presently accommodates ~20 mm/yr of strike-slip and forms the main linkage structure between the Alpine Fault and the Hikurangi subduction zone [Van Dissen and Yeats, 1991; Langridge et al., 2003].
1.4.2 Basement geology

No Precambrian cratonic core is exposed in New Zealand (GNS Science, 2012). The basement rocks in New Zealand consist of nine Cambrian to Early Cretaceous volcani-sedimentary tectonostratigraphic accreted terranes related to its history as a convergent margin of the Gondwana supercontinent (Figure 1.2) [e.g., Mortimer, 2004]. Three regional batholiths (e.g., Median Tectonic Zone) and three regional metamorphic belts (e.g., Haast Schist) overprint the terranes [Mortimer, 2004]. Late Cretaceous sea floor spreading rifted Zealandia from the landmasses of Australia and Antarctica to open the Tasman Sea [e.g., Gaina et al., 1998; Sutherland, 1999a]. Tertiary sedimentary rocks and volcanic cover commonly

![Figure 1.2](image.jpg)

Figure 1.2 Map of South Island Basement Terranes. Main Cambrian-Cretaceous tectonostratigraphic terranes (and cross-cutting batholiths and belts of regionally metamorphosed rock) of the South Island of New Zealand simplified after Mortimer [2004]. Note ~460 km dextral offset of the Dun Mountain-Maitai terrane across the Alpine Fault and bending of the basement terranes. Refer to Figure 2.1 for lithologic constituents of terranes in the study area. Abbreviation used: E (Early), M (Middle), L (Late), C (Cambrian), O (Ordovician), D (Devonian), P (Permian), Tr (Triassic), J (Jurassic), K (Cretaceous).
obscure comparable underlying basement terranes on the North Island and most of the terranes are best exposed in the South Island (Figure 1.2). A distinctive belt of ultramafic rocks (a portion of the Dun Mountain-Maitai terrane) exposed in the Nelson area (northeast South Island) and in South Westland (southwest South Island) on opposing sides of the Alpine Fault led Wellman to propose a once controversial ~480 km strike-slip offset across the Alpine Fault [Wellman, 1949 quoted in Benson, 1952]. Although actually closer to 460 km, this is now widely accepted as the minimum Neogene offset of the Alpine Fault. This is a minimum displacement because it does not account for displacement partitioned off of the northern Alpine Fault and onto the Marlborough Fault System or distributed crustal shear recorded in the Cenozoic bending of the basement terranes (Figure 1.2) [e.g., Sutherland, 1999b].

1.4.3 South Island geosystem

Obliquely convergent plate boundary motion in the South Island has driven complex feedback between erosional and depositional processes that denude or decrease relief (e.g., glaciations, rainfall, erosion; Figure 1.3) and geological drivers which create relief in the island-wide, tectonically-active Southern Alps/Fiordland orogen (Figure 1.4) [Cox et al., 2012a and references therein]. The mountain belt presents a topographic barrier that results in enhanced rainfall from the moist westerly winds to produce a very asymmetric pattern of rainfall and erosion west and east of the main divide; high crustal heat flow and advection of crustal isotherms associated with the resultant exhumation causes thermal weakening of the Alpine Fault zone at depth, which helps focus oblique slip on this single structure [Koons et al., 2003]. During the last 5 Ma of oblique convergence, the Southern Alps Orogen has achieved high exhumation rates of ~6–9 mm/yr in the central section and a total convergence > ~40 km [Batt et al., 2000; Little et al., 2005]. This has produced a crustal root beneath the orogen extending a further ~10–20 km deeper than adjacent crust, which comprises a delaminated lower crust and a thickened overlying middle crust [Davey et al., 2007]. Despite a historical record of low levels of seismicity on the Alpine Fault [e.g., Leitner et al., 2001] and no evidence of fault creep [Sutherland et al., 2007a and references therein], there is abundant paleoseismological evidence that the fault ruptures quasi-periodically in large to great magnitude earthquakes (~M 8.0) and is the single most important driver of geomorphology on the South Island (Figure 1.5) [e.g., Wells and Goff, 2006, 2007; Sutherland et al., 2007a and references therein; Berryman et al., 2012a; Howarth et al., 2012; this study].
1.5 Fieldwork

The remoteness and climate of the region present the greatest logistical challenges to fieldwork. Fieldwork consisted of a total of 76 field days or about three months including travel days. A complete set of stereoscopic aerial photographs and Google Earth satellite imagery of the area were employed to support geomorphic observations and to plan fieldwork. A total of 737 field localities were georeferenced in the course of study, typically to ±10 m accuracy in the bush (Figure 1.6).
1.5.1 Physiography

The Alpine Fault forms the main physiographic feature of the field area, separating the 2000–3000 m peaks of the main divide of the Southern Alps to the southeast from low gradient
Figure 1.5  South Island Geomorphic Provinces. Major tectonically-induced geomorphic provinces of the South Island of New Zealand [this study, after Williams, 1991] overlain on a slope map (high slopes in white, low slopes in black) derived from LINZ data. Regions were defined based on fault and fold style, as well as geomorphology (utilizing maps of topography and slope). The present study focuses primarily on the Westland Coastal Plains and Hills, Southern Alps, and Fiordland geomorphic provinces nearest the southern onshore half of the Alpine Fault.

river valleys and circular island-like bedrock ranges to 1100 m elevation northwest of the fault. Treeline is typically at ~1100 m elevation, but is as low as 600 m in the presence of iron-rich ultramafic rocks. Major glaciated river valleys have their headwaters in the Southern Alps and drain northwest to the Tasman Sea unless deflected by the Alpine Fault. Rivers and steep creeks with headwaters northwest of the Alpine Fault typically have eroding bedrock landslips at their heads and display little to no evidence of former glaciations. Wide river valleys or low elevation saddles along the Alpine Fault are typically poorly drained and lowland swamps are common. Lakes McKerrow, Alabaster and Wilmot in the Hollyford and Pyke valleys form significant features less than 30 m in elevation (Lake McKerrow is tidally

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Figure 1.6 Main Field Area. Overview of the main field area (see Figure 1.1) with means of logistical access and georeferenced field localities utilized in this study (note: more localities were visited than were georeferenced). Northwest-illuminated hillshade is derived from 15 m LINZ DEM data and overlain on 75 m bathymetry data courtesy of NIWA.

Influenced). The Hollyford Valley marks the boundary between the Southern Alps to the northeast and Fiordland to the southwest; the latter is an area of steep U-shaped valleys and glacial fiords. The coast is often boulder-lined, but with beaches at the mouths of the major rivers and cliffs plunging into the sea in places.
1.5.2 Environment

Prevailing westerly winds account for mean annual rainfall in excess of 10 m (Figure 1.3) along the western rangefront of the Southern Alps, with rain occurring on about half of the days of a year. A successful week of fieldwork on the West Coast (carefully planned using long range weather forecasts) frequently consists of about 5 low precipitation days and two moderate to high precipitation days during which productivity can be limited. This high rainfall supports a dense temperate rainforest with a canopy of beech and/or podocarp and an understory of ferns which can be thick enough to obscure the ground, the former of which effectively block the sun and hinder GPS signals. In steep terrain, these ferns can provide essential hand and foot holds. Moss, mud and decaying logs typically coat the forest floor.

Because of New Zealand’s unique tectonic history, it has been fortunate (for field geologists) to evolve flora and fauna with a relatively benign effect on humans. The native variety of stinging nettle (ongaonga) and thorned vines (e.g., bush lawyer) are generally easily avoided. Perhaps of greatest nuisance are the small biting insects known as sandflies. They are ubiquitous along the coast and lowlands and can form thick clouds that target exposed skin and form a sort of psychological warfare which hinders thought and productivity. They are worst in summer and comparatively benign in the winter. New Zealand has no poisonous snakes, scorpions or spiders (excepting a single rare species). Birds have evolved to fit every niche up to top predator with two species of bats being the only native land mammals. On the coast fur seals are best avoided during breeding season (mid-November – mid-January). Introduced mammal species are limited to stoats, possums and deer. Deer are notable for creating anastomosing trail networks which are generally the best means of travel through the bush. They also account for one of the most dangerous hazards: man, or more specifically, hunters. During hunting season (March – August) and especially during deer breeding season (late March – late April), it is best to notify intentions and wear a high-visibility vest. It is very uncommon to see other people away from roads in the field area, but this means they too will not be expecting to see other people.

Mean annual temperatures in the region range from about 8–11 °C at low elevations with typical annual highs around 25 °C and sub-zero lows. Snow can fall at any time of year. The weather can change rapidly and rivers can rise fast. Most rivers in the study area are
unbridged and great care is required to safely cross them. Some rivers are only able to be crossed at a handful of locations in normal conditions.

1.5.3 Exposure

South Westland is densely-vegetated and outcrop often exists as a serendipitous and transient exception to the norm. Outcrop quality varies from excellent to poor and commonly occurs in a few key contexts (Figure 1.7). The most important outcrops in this study are those associated with steep, bedrock-eroding creeks, especially when a recent flood event has cleared the bedload or forced the creek to migrate laterally. The meters above bedrock just above creek level (when water levels are low) are typically stripped of moss and other vegetation and provide superb outcrop that is sometimes continuous for > 100 m (Figure 1.7A). The best Alpine Fault outcrops are those associated with a steep creek flowing orthogonal to fault strike. Fresh landslips also generally provide important (albeit highly-fractured) outcrop; these too become overgrown with vegetation on relatively short timescales (Figure 1.7B). Outcrops in alpine environments are good though commonly frost-shattered and heaved, which adds a degree of uncertainty to structural measurements (Figure 1.7C). Outcrop quality can also be very good in coastal cliffs, sea stacks and on shore platforms (Figure 1.7D). Occasionally a recently uprooted tree can provide a fortuitous outcrop in heavily-bushed areas. Conditions of low water levels are generally preferable as they reveal outcrop on creek, river and lake shores obscured at higher flows.

Many of the hydrothermally-altered, clay-rich and highly-fractured fault rock samples examined in this study have received limited scientific attention until recently, in part due to their generally poor exposure elsewhere on the Alpine Fault and partly due to the difficulty of sampling and thin sectioning these materials. Collected samples were always oriented in the field by marking with a permanent marker or a grease pencil if wet. If dry, densely-fractured samples were glued in situ before collection. However, these rocks are rarely dry in the field, so more commonly a sample was carefully removed and immediately wrapped in aluminum foil to be thoroughly coated in glue back at the laboratory (while still partly in its foil jacket). After gluing, fractured samples could be prepared as normal thin sections. Great care was necessary to extract intact clay gouge samples for thin sections and friction experiments; the most successful method employed a wide bricklayer’s chisel which could be driven into the gouge and used as a shield while excavating the material on the opposite side as the sample.
Figure 1.7 Outcrop Types. (A) Typical creek section highlighting the good quality of outcrop within a few meters of the water surface (above this outcrop is obscured by moss and other vegetation; Durwards Creek area), (B) example of exposure in an actively-eroding slip (good, but commonly highly-fractured; Kaipo Slips), (C) alpine exposure showing frost-shattered and lichenated rocks (Telescope Hill), and (D) coastal outcrop in shore platforms and sea stacks (Gorge River mouth).
Considerable time and effort was spent developing a technique to produce high quality thin sections of clastic clay-rich gouges (procedure detailed in the Appendix).

1.5.4 Logistics and methods

Road access to the field area is limited to the Jackson River Road in the north and the Hollyford Valley Road in the south (Figure 1.6). Hiking between these road ends at a swift pace (no fieldwork) with a light pack (no rocks) takes about 8 days. Many areas are not feasible to access from these road ends and require motorized transport (helicopter, fixed-wing aircraft, jet boat) to make effective use of time and weight limitations (Figure 1.8). I was fortunate to be only a half day’s drive away from the field area meaning that I could place field assistants and pilots on standby and wait on a good weather forecast to maximize the productivity of the intended fieldwork. This method is recommended for anyone working in the area.

Fieldwork was done year round; summer has long days, warm weather but sandflies, while winter has more settled weather, low river levels, few sandflies, but short and cold days. Other than day trips from road ends and backpacking trips up to 8 days in length, trips were often of a “fly-in, fly-out” nature with day trips away from a camp. Helicopters frequently operate out of Milford Sound and Neils Beach (Jackson Bay), fixed-wing aircraft out of Milford Sound, and jet boats in the Lower Hollyford valley (the latter only during tourist season). Helicopters provided the most versatile mode of transport, often being able to deliver us right to a desired basecamp (Figure 1.8A). Fixed-wing aircraft were a cost-effective solution but required better weather and depending on the site, appropriate tides (Figure 1.8B). Jet boats were used for transport across Lake McKerrow (Figure 1.8C).

Several novel approaches were utilized to access key outcrops including inflatable boat (Figure 1.8D), single rope technique abseiling (Figure 1.8E) and canyoning (Figure 1.8F) to great success. Crossing rivers often proved a formidable challenge (Figure 1.8G). Below bushline flat ground is often swampy and suitable campsites were few and far between (Figure 1.8H). For safety a personal locator beacon (EPIRB), mountain radio or satellite phone was carried at all times. Map, compass and GPS were used for navigation and at field localities.
Figure 1.8  Fieldwork Logistics. (A) Other than walking, helicopters provide the most versatile transport in the field area and are able to land in small clearings and on uneven ground (Wolf River). (B) Fixed-wing aircraft is a more cost-effective transport option that is able to land on a small number of airstrips or certain beaches at low tides (Big Bay). (C) Jet boats are useful for crossing Lake McKerrow in the Hollyford valley when operators are present in the high tourist season (Pyke-Hollyford confluence). (D) An inflatable raft was used to access key outcrops in important sections (Martyr River). (E) Important outcrops in slips and waterfalls were measured and collected using single rope abseiling techniques (Saddle Creek South). (F) Canyoning techniques were used to access undescribed outcrops in Monkey Puzzle Gorge. (G) Crossing rivers is often a significant consideration; here a serendipitously-placed log provided the only safe crossing of the Kaipo River within a > 2 km stretch of river rapids. (H) Setting up a typical campsite in the McKenzie Creek region. This site, just big enough for a single tent, was the only flat and well-drained ground seen along ~2 km of the creek.
Place names and physiographic features referred to in this thesis are those found on published Land Information New Zealand (LINZ) 1:50,000 scale Topo50 maps. At present these maps can be freely viewed or downloaded from the LINZ website. Italicized names used are unofficial names introduced by this thesis (e.g., Madagascar Creek). Where applicable they refer to nearby landmarks or the nature of the feature, otherwise they commemorate field assistants. Pertinent petrologic samples and thin sections are lodged in the University of Otago, Department of Geology’s research collection (given as OU#) and can be correlated to field samples utilizing the table included in the Appendix. Field sample numbers are of the form N-year-day-month-letter (e.g., N100102A, N100102I). A PDF document containing scans of all PhD field notes is provided on the DVD, as well as a GIS-readable Excel file. Fossil collections assembled in this study were entered into the New Zealand Fossil Record Electronic Database (FRED) which can be accessed online; the fossil collections pertaining to this thesis are lodged in the University of Otago Department of Geology research collection. New formations names introduced by this study (e.g., Kaipo Mélange, Sara Formation, Wolf Formation) have been submitted to the New Zealand Stratigraphic Lexicon (StratLex), also available online.

1.6 Publications and Conferences

Several conferences and workshops were attended during the course of the PhD at which thesis research was presented. A full list of abstracts from conferences is available in the Appendix. The papers and manuscripts listed below are modified versions of the indicated chapters of this thesis:


southern Alpine Fault and implications for the New Zealand plate boundary. *Journal of Structural Geology* special issue “Continental transpressive fault zones” (Chapter 5)

### 1.7 Acknowledgement of Contributions

The diverse nature of this PhD necessitated numerous collaborations; in addition to utilizing resources and facilities available at the University of Otago, collaborations were made with peers at the University of Canterbury, GNS Science, University of Western Australia, University of Wollongong, Pennsylvania State University, University of Liverpool and Universität Bremen to make use of knowledge and facilities not present at Otago. The breadth and impact of this thesis would have been much reduced without their contributions. The data, results and interpretations present in this thesis are my own unless indicated below or as cited in the text.

**Chapter 2:** Samples for x-ray diffraction (XRD) and friction/permeability experiments utilized in this study were collected by myself with the exception of the Gaunt Creek samples, which were collected by collaborator Carolyn Boulton (University of Canterbury). Carolyn prepared samples for XRD analyses by the Commonwealth Scientific and Industrial Organisation (CSIRO) and performed the friction/permeability experiments in conjunction with Brett Carpenter (Pennsylvania State University). The XRD and friction/permeability data presented in this chapter are also discussed as part of a larger suite of samples collected and analyzed by Carolyn for her PhD thesis. Carolyn provided text on friction/permeability experimental methods which has been largely unchanged. Carolyn created Figure 2.7 and performed the friction rate parameter (a-b) and frictional healing (Δμ) calculations therein. While ultimately written myself, the text of this chapter was enhanced by Carolyn’s suggestions. Prepared samples were sent to Geoffrey Batt (University of Western Australia) for 40Ar/39Ar analyses. Geoffrey provided a summary of the experimental procedures. Collaborators Carolyn Boulton, Brett Carpenter, Geoffrey Batt and supervisor Virginia Toy (University of Otago) provided feedback which influenced my interpretations.

**Supplement 2:** The interferometric synthetic aperture radar (InSAR) study was carried out collaboratively with Isabelle Ryder (University of Liverpool). I selected the area and imagery to use and Isabelle processed the data; the interpretations of the data are my own. Unpublished results of the Lake McKerrow triangulation and trilateration survey were provided by John Beavan (GNS Science). Focal mechanism and P-axis solution data for earthquakes in the study area were provided by John Ristau (GNS Science).
Chapter 3: Supervisor Virginia Toy performed the analyses of fault kinematics, stability and slip tendency utilizing my suggestions of fault orientations to input. The corresponding sections detailing these analyses were largely written by Virginia and she created Figures 3.7 and 3.8. The results of these analyses were arrived at collaboratively. Supervisors Virginia Toy and Richard Norris (University of Otago), and collaborator Rob Langridge (GNS Science) provided feedback which influenced my interpretations.

Chapter 4: Both radiocarbon dates utilized (NZ 4626; WK4914) have been previously published as indicated. Dawn Chambers (GNS Science) provided the original records for NZ 4626 for my use. Tim Davies (University of Canterbury) provided feedback on the manuscript content which influenced my interpretations, and provided suggestions on writing style.

Chapter 5: Colin Murray-Wallace (The University of Wollongong) analyzed the fossil shells using amino acid racemization. Alan Beu (GNS Science; macropaleontologist), Bruce Hayward (Geomarine Research; micropaleontologist), Denise Kulhanek (GNS Science; nannopaleontologist) and Dallas Mildenhall (GNS Science; palynologist) provided identifications and interpretations (water depth, environment, age, etc.) according to their specialties.

Supplement 5: Colin Murray-Wallace (The University of Wollongong) provided the analytical methods section on amino acid racemization dating.

Appendix: Paleontological reports presented are by the authors as indicated. These data and interpretations have been incorporated into my own interpretations in Chapter 5.
Chapter 2  Nature and Timing of Slip Localization on the Southern Alpine Fault, New Zealand

Figure 2.0  Oblique aerial photograph looking northeast over Lake McKerrow, along the Alpine Fault (vertical at center). At center is the alluvial fan of Hokuri Creek. Hokuri Creek is the site of a c. 8000 yr near-fault paleoseismic record of large magnitude Alpine Fault surface-rupturing earthquakes displaying remarkably quasi-periodic behavior [Berryman et al., 2012a]. At this same location, the present study describes a 12 m-wide Alpine Fault fault core composed of saponite-chlorite-lizardite fault gouge with large (m-scale) serpentine clasts. These frictionally weak gouges (the weakest yet observed on the Alpine Fault) have been identified in outcrop for ~10 km along-strike, associated with an Australian plate-hosted tectonic mélangé identified in this study (Kaipo Mélange), which extends at least 40 km along-strike of the Alpine Fault. The results of this study highlight a conundrum: wide fault cores containing weak fault rocks with textures and frictional properties suggestive of slow aseismic slip are exposed, despite abundant evidence that large magnitude earthquakes occur here. Width of Lake McKerrow along the Alpine Fault is ~2.5 km. Photo taken by L. Homer (CN6212/1 H). Photo courtesy of GNS Science.

This chapter is a version of a submitted manuscript: Barth, N.C., C.J. Boulton, B.M. Carpenter, G.E. Batt, and V.G. Toy. Slip localization on the southern Alpine Fault, New Zealand. Submitted to Tectonics. Details of the various author contributions are given in Chapter 1. See also Supplement 2 for additional data and discussions.
Abstract

Results of a detailed field study of the southern onshore portion of New Zealand’s Alpine Fault reveal that for 75 km along-strike, dextral-normal slip on this long-lived structure is highly-localized in phyllosilicate-rich fault core gouges and along their contact with more competent rocks. At three localities (Martyr River, McKenzie Creek, Hokuri Creek), I document complete cross sections through the fault. New $^{40}\text{Ar}/^{39}\text{Ar}$ dates on mylonites, combined with microstructural and mechanical data on phyllosilicate-rich fault core gouges show that modern slip is localized onto a single, steeply-dipping 1 to 12 m-thick fault core composed of impermeable ($k = 10^{-20} - 10^{-22}$ m$^2$), frictionally weak ($\mu_s = 0.12 - 0.37$), velocity-strengthening, illite-chlorite and saponite-chlorite-lizardite fault gouges. Fault core materials are (1) comparable to those of other major weak-cored faults (e.g., San Andreas Fault), and (2) most compatible with fault creep, despite paleoseismic evidence of quasi-periodic large magnitude earthquakes ($M_w > 7$) on this portion of the Alpine Fault. I conclude that frictional properties of gouges at the surface do not characterize the overall seismogenic behavior of the southern Alpine Fault.

2.1 Introduction

A primary objective of earthquake science research remains understanding the factors that influence variations in seismological behavior, and achieving this objective requires integrating data on the structure and mechanical behavior of faults [e.g., Segall, 2012; Noda and Lapusta, 2013]. Because of their relatively large fault plane area and geometrical simplicity, large-displacement, crustal-scale faults represent globally significant seismic hazards. On these structures, strain is thought to localize as quasi-static and/or dynamic wear processes in the brittle crust gradually eliminate structural complexity with increasing displacement [Segall and Pollard, 1983; Wesnousky, 1988; Ben-Zion and Sammis, 2003; Biegel and Sammis, 2004]. At all crustal depths, including below the brittle-ductile transition, these processes are accompanied by strain softening through grain size reduction, recrystallization, reaction softening, alignment of weak material components forming a pervasive fabric, shear zone coalescence and shear heating [Wintsch et al., 1995; Goodwin and Wenk, 1995; Handy and Brun, 2004; Holdsworth, 2004; Jefferies et al., 2006].

With a remarkably straight surface trace striking northeast-southwest for over 800 km, and a cumulative displacement of $> 460$ km, the Alpine Fault is a continental transform fault that accommodates about 70% of the relative motion between the Australian (AUS) and Pacific
(PAC) plates in the central South Island of New Zealand [Norris and Cooper, 2001]. It represents both a mature crustal-scale plate boundary fault and the largest onshore seismic hazard in New Zealand [Sutherland et al., 2007]. Since c. 45 Ma, relative plate boundary motion has resulted in various modes of deformation accommodated by faulting through the New Zealand continent (i.e., Zealandia), beginning with asymmetric rifting in the late Eocene (c. 45–35 Ma), followed by transtension from the late Eocene to Oligocene (c. 35–25 Ma), wrench faulting from the early to late Miocene (c. 25–6 Ma), and oblique strike-slip motion since the late Miocene (c. 6 Ma) [Kamp, 1986; Cooper et al., 1987; Walcott, 1998; Sutherland et al., 2000; Cande and Stock, 2004; Reyners et al., 2011]. Modern GPS data coupled with geologically-based estimates over the last 3 Myr yield average relative plate velocities of ~40 mm/yr, with 9 ± 1.5 mm/yr normal to, and 38.9 ± 1.0 mm/yr parallel to the plate boundary [Beavan et al., 2002; DeMets et al., 1994; 2012; Figure 1.4D].

With a last known surface rupturing earthquake that occurred in 1717 AD [Wells et al., 1999], an estimated mean recurrence interval of 329 ± 68 years [Berryman et al., 2012a], and average slip rates between 20–30 mm/yr [Norris and Cooper, 2001], the Alpine Fault is late in its seismic cycle. Since 2004, the San Andreas Fault Observatory at Depth (SAFOD) project has greatly advanced our understanding of fundamental processes of rock deformation in fault zones and earthquake physics [e.g., Zoback et al., 2010]. A similar project, the Deep Fault Drilling Project (DFDP), located on the central Alpine Fault, is now underway with the goal of sampling and monitoring the fault to a few kilometers depth [Townend et al., 2009]. Results from the first phase of drilling [Sutherland et al., 2012], together with recent light detection and ranging (LiDAR) data [Barth et al., 2012; Chapter 3], paleoseismicity studies [e.g., De Pascale and Langridge, 2012; Howarth et al., 2012], laboratory measurements of fault rock frictional and hydrologic properties [Boulton et al., 2012], and seismicity catalogs from the Southern Alps Microearthquake Borehole Array (SAMBA) [Boese et al., 2012; Wech et al., 2012], are now rapidly augmenting our knowledge of the central Alpine Fault.

Comparatively, the southern onshore Alpine Fault (hereafter termed the southern Alpine Fault) is much less studied. Seismicity catalogs, geological mapping, and tectonic reconstructions indicate that the southern Alpine Fault has structural, seismological and lithological properties distinct from the central section [e.g., Berryman et al., 1992; Sutherland et al., 2000; GNS Science, 2010; Norris and Cooper, 2007; Reyners et al., 2011; Geonet, 2012]. In this chapter, using $^{40}$Ar/$^{39}$Ar dating, I provide new ages for formation of mylonites in both the footwall and hanging wall, which show the southern Alpine Fault exploits a Cretaceous zone.
of shear and that the Alpine Fault accommodated displacement in the modern plate boundary configuration since at least the Early Miocene. I also document the first three complete cross sections through the fault, which in places is composed of the Mg-rich phyllosilicate mineral saponite, a weak mineral widely cited as the cause of fault creep on the San Andreas Fault [Carpenter et al., 2011; Holdsworth et al., 2011; Lockner et al., 2011; Moore and Rymer, 2012].

2.2 Tectonic Setting

2.2.1 Geologic history

New Zealand comprises the emergent portion of a mostly submerged continental landmass, Zealandia. No Precambrian cratonic core is exposed in New Zealand [GNS Science, 2012], and basement rocks consist of nine Cambrian to Early Cretaceous volcani-sedimentary tectonostratigraphic terranes accreted to Zealandia when it comprised the convergent margin of the Gondwana supercontinent [e.g., Mortimer, 2004 and references therein; Figure 1.2]. In the South Island, the basement terranes are classified into the Cambrian-Devonian Western Province and the Permian-Cretaceous Eastern Province, stitched together by the Carboniferous-Early Cretaceous Median Batholith [e.g., Bishop et al., 1985; Mortimer et al., 1999].

Late Cretaceous sea floor spreading rifted Zealandia from the continental masses of Australia and Antarctica to create the Tasman Sea [e.g., Gaina et al., 1998; Sutherland, 1999]. A distinctive belt of ultramafic rocks has been offset 460 km by the Alpine Fault in the Neogene (Figure 2.1A). While spatially much of the South Island is dominated by pelitic greyschist/greenschist and its sedimentary protoliths (purple on Figure 2.1), there is a high lithologic diversity represented by the terranes overall (Figure 2.1). This lithologic diversity is highest in the study area.

Figure 2.1 Geologic Setting. (A) Simplified basement lithology map of the South Island of New Zealand with major plate boundary structures. Alpine Fault sections are those defined in Table 2.1. Mapping after GNS Science 1:1,000,000 geological map [2009]. Plate motion vectors from NUVEL-1A [DeMets et al., 1994]. Dotted line denotes the continuation of the Permian Dun Mountain Ophiolite Belt (DMOB), which has been dextrally offset 460 km across the Alpine Fault [Wellman, 1955]. (B) Simplified basement lithology map of the study area showing relative heterogeneity of rock types adjacent to the Alpine Fault overlain on LINZ 15 m DEM. A basement lithology map is presented rather than a more traditional basement terrane map to emphasize the diversity of rock types adjacent to the southern Alpine Fault and to point out likely correlations of rock types across major structures (e.g., the Glade-Darrans Fault). Lithologies correspond roughly to Permo-Triassic basement terranes (Figure 1.2). Both active and inactive faults are shown. Mapping after Rattenbury et al. [2010],
Turnbull et al. [2010], Barnes et al. [2005], Turnbull [2000], and this study. Abbreviations and correlations: MA (Martyr River), MC (McKenzie Creek), H (Hokuri Creek), SS (Sutherland Sound), ASZ (Anita Shear Zone), BSVG (Brook Street Volcanic Group; Brook Street terrane), DMUG (Dun Mountain Ultramafics Group; Dun Mountain-Maitai terrane, synonymous with the Dun Mountain Ophiolite Belt), Greenland Group (Buller terrane), Haast Schist (Torlesse composite terrane, Rakaia terrane), LVG (Livingstone Volcanics Group; Dun Mountain-Maitai terrane), MWG (Mt Webb Gneiss; correlation uncertain, likely Paleozoic protolith of an Anita Shear Zone lithology).

2.2.2 Along-strike variations in fault structure

Here I focus on improving our understanding of the southern Alpine Fault, formally named the South Westland section. There are distinct along strike differences in fault properties
which allow subdivision into sections, as proposed by Evison [1971], Berryman et al. [1992], Bull [1996], Barnes et al. [2005], and Sutherland et al. [2007]. Distinguishing structural, seismic and geomorphic features of these major sections of the fault are summarized in Table 2.1. I place the northern boundary of the southern Alpine Fault, the focus of this contribution, at the Martyr River where there is an abrupt change in uplift polarity (discussed in Section 2.3.1) and the southern boundary at Caswell Sound where there is a 4 km-wide step-over in the fault plane (Figure 2.1A). The Central section of the Alpine Fault has long been recognized as having low background seismicity to as far south as ~25 km north of the Martyr River [e.g., Evison, 1971]. In contrast, the southern Alpine Fault has higher rates of post-1940 recorded seismicity [Evison, 1971; Anderson and Webb, 1994; Wallace et al., 2007; Boese et al., 2012; GeoNet, 2012].

Offshore Jackson Bay marks the transition from a continental/continental transpressive margin to a continental/oceanic transpressive margin in the south; this is the only such transition on the fault (Figure 2.1B) [Berryman et al., 1992]. Outboard of the continental shelf southwest of Jackson Bay, a youthful, north-propagating subduction zone subducts the AUS plate obliquely beneath the Alpine Fault [Barnes et al., 2002]. The associated Wadati-Benioff zone extends as far north as Hokuri Creek in the Hollyford Valley (Figure 2.1B) [GeoNet, 2012]. An unfaulted onshore AUS plate block provides a stark contrast to lithologically diverse Paleozoic-Mesozoic basement terranes on the PAC plate where both inherited and newly formed structures are present (Figure 2.1B). In addition to the Alpine Fault, major faults in the area include the Pembroke, Glade-Darrans, Hollyford, Livingstone, Greenstone and Moonlight fault systems (Figure 2.1B). While the geology in this region is generally well-mapped, large uncertainties persist about the extent to which structures have recently been active [cf. Cox et al., 2012].

Previous studies of the southern Alpine Fault have largely been at the regional scale and reconnaissance level [Wellman and Willet, 1942; Clark and Wellman, 1959; Wellman and Wilson, 1964; Hull and Berryman, 1986; Berryman et al., 1992; Campbell, 2005]. Sutherland and Norris [1995] focused on slip rates and tectonic geomorphology, but included observations of the structure of the Alpine Fault zone at Hokuri Creek where they noted altered ultramafic blocks within a wide zone of clay-rich fault gouge. Also at Hokuri Creek, Berryman et al. [2012a] used an exposed sequence of silt/peat cycles ponded against the Alpine Fault scarp to obtain a quasi-periodic record of earthquake events over the last 8000 years. Berryman et al. [2012a] infer that the event record at this site reflects the fault’s geometrically simple
structure, high slip-rate, and tendency to work in isolation from other faults. In this chapter, I use field observations and lab-based analytical techniques to describe and quantify the architecture, mechanical behavior, and tectonic history of the southern Alpine Fault.

2.3 Field Observations

2.3.1 Plate boundary structure and kinematics

Surface outcrops of the southern Alpine Fault were mapped at 10 major locations along-strike. In this remote, rainforest-covered region, the fault core is well-exposed at five locations. Slickenlines on fault planes at these locations are consistent with the sense of offset indicated by widespread fault scarp geomorphic observations. An abrupt change in kinematics and geomorphic expression is located immediately southwest of the Martyr River without a major change in fault strike, fault dip or jogging of the outcropping fault plane (Figure 2.2). To the northeast on the central Alpine Fault, fault kinematic data indicate oblique motion is accommodated on a 055°-striking, ~45° southeast-dipping dextral-reverse fault plane causing net uplift of the PAC plate. To the southwest on the southern Alpine Fault, the fault plane steepens to 80–90° southeast while maintaining a ~052° strike; here the recent net uplift is of the AUS plate despite the overall higher elevation of ranges east of the fault. Though the motion resolved on the Alpine Fault here is almost purely strike-slip, a ubiquitous normal component is associated with net uplift of the AUS plate (Figure 2.2). This relationship appears to continue for at least 75 km offshore to Caswell Sound [Barnes et al., 2005]. On the southern Alpine Fault, steep northwest-draining creeks cut orthogonally to the Alpine Fault, and provide the best outcrops of both sides of the plate boundary structure through the entire width of the damage zone.

2.3.2 Fault zone architecture

In this study, fault zone architecture is described using the simple fault core zone and damage zone model proposed by Caine et al. [1996], which Sutherland et al. [2012] adopted for the central Alpine Fault. I define the damage zone as the tabular zone in which minor faults, folds, veins and fractures have formed because of deformation associated with faulting. Within this damage zone, there is a fault core that contains fault rocks formed by particle size-reduction in the brittle regime and dynamic recrystallization of highly strained grains in the ductile regime [Reed, 1964; Sibson, 1977; Sibson et al., 1981; Norris and Cooper, 2007]. The fault core (or cores) may contain one or more principal slip surfaces (PSSs) which
Table 2.1

Alpine Fault Sections

Section name
Informal name
Length (km)
Northeastern extent
Southwestern extent
Strike-slip rate (mm/yr)
NUVEL-1A plate vector (mm/yr→°)
Current % of plate boundary motion
Regional fault strike (> 10 km lengths)
Section average fault strike/dip
Dip-slip rate (mm/yr)
Net uplifted side
Along-strike lithological variation
Single event strike-slip displacement (m)
Single event dip-slip displacement (m)
Locking Depth from GPS (km)
Depth of base of seismogenic zone (km)
Background seismicity
Intersections with other faults
Tectonic complexities
Major transitions
Large scale fault bends
Wairau
Wairau Fault
200
Cook Strait
Matakitaki River
3 – 6.7
39.6 → 258°
8 – 17%
056° – 081°
067°/90
~0
x
Low
5–7
0
20–30
10 – 12
Very Low
Few
MFS rotation
NIFS
RSB & LSB OFF

1
North Westland
northern Alpine Fault
150
Matakitaki River
Toaroha River
10 – 13.6 (-2/+1.8)
38.4 → 255°
26 – 35%
006° – 055°
055°/55°SE
3.4 – 6 (-0.6/+2)
SE
Generally low
~6
≤3
~12
10 – 12
Low
Many
MFS interaction
MFS
20 km wide LSB

2

Central
central Alpine Fault
250
Toaroha River
Martyr River
27 – 29 (-5/+6)
37.5 → 251°
72 – 77%
052° – 060°
055°/45°SE
2.25 – 8 (-0.5/+1)
SE
Low
8–9
1–<4
13–18
8–9
Low
Few
SP (< 2km depth)
OM FR
None

3

South Westland
southern Alpine Fault
160
Martyr River
Caswell Sound
23 – 27.2 (-2/+1.8)
36.6 → 247°
63 – 74%
040° – 059°
052°/82°SE
0.6 (-0.2/+0.3)
NW
High
7.5 – 9 (-0.5)
~1
Unknown
10 – 12
High
Many
DS at CR
TF to SZ, CC to OC
10° B ON, 20° B OFF

4

Fiordland
Resolution section
130
Caswell Sound
West of Dusky Sound
31.4 (-3.5/+1.8)
35.9 → 244°
87%
036° – 056°
040°/75°SE
~0
NW
Low
Unknown
Unknown
Unknown
10 – 12
High
Moderate
AW, Rooted into OM SZ
Large SO
3 major LSBs

5

AW: accretionary wedge
B: bend (in fault strike)
CC: continental crust
DS: duplex structure
FR: fault ramping
LSB: left stepping bend
MFS: Marlborough Fault System
NIFS: North Island Fault System
OC: oceanic crust
OFF: offshore
OM: oblique motion
ON: onshore
RSB: right-stepping bend
SP: serial partitioning
SZ: subduction zone
TF: transform fault

30

30


accommodated a significant portion of the total offset across the fault zone [Sibson, 2003]. I also use alteration zone to describe the region where fluid-rock interactions alter the physical, chemical and hydrological properties of the fault zone [Sutherland et al., 2012]. On the Alpine Fault, fluid-rock interaction within the alteration zone occurred primarily at greenschist facies temperatures and pressures, as indicated by the modally dominant chlorite and epidote alteration minerals [e.g., Warr and Cox, 2001; Sutherland et al., 2012]. Within this zone, carbonate vein abundance is also above regional levels [Warr and Cox, 2001; Vry et al., 2001; Norris and Cooper, 2007].

While fault cores commonly comprise thin (mm to cm-wide) planes [Ben-Zion and Sammis, 2003; Sibson, 2003], wider, branching, anastomosing fault cores are also common worldwide [Faulkner et al., 2003; 2010]. The evolution of fault strength during coseismic slip depends critically on the width of the actively slipping portion of the fault core [Marone and Kilgore, 1993], and field observations of active fault cores provide information necessary for numerical modeling of earthquake processes [e.g., Rice and Cocco, 2006; Niemeijer et al., 2012].

I present detailed cross sections of the Alpine Fault core and damage zone at three localities. Fault rocks described are classified into protoliths, damaged rocks including protocataclasites, and fault core rocks including random fabric cataclasites, foliated cataclasites, and fault gouges. Fault rocks are classified based on the scheme proposed by Sibson [1977] and modified to include the presence of foliated cataclasites with an interconnected network of
phylllosilicate minerals [Holdsworth, 2004; Jefferies et al., 2006]. A separate type of fault rock, pulverized rock, characterized by a fracture-induced reduction in grain size with minimal distortion of the primary rock structure [Dor et al., 2006, 2009; Wilson et al., 2005], was not observed in the study area.

Figure 2.2  Alpine Fault Kinematics. (A) Strip map of the South Westland section of the Alpine Fault (area shown in Figure 2.1). Location names refer to creeks or rivers labeled. Up/Down (U/D) and strike-slip motion is constrained by geomorphology. Offshore description is from Barnes et al. [2005]. Stars denote localities shown in Figure 2.5. (B) Equal area lower hemisphere stereonet with best representative measurements of slickenlines (circles) on Australian-Pacific principal slip surface (great circles and poles to planes as squares) at locations shown in A. Variability of strike, dip, trend, and plunge at each location is typically < 5–10° and all measurements are taken at or close to where the fault juxtaposes bedrock on bedrock. John O’Groats measurements are from Clark and Wellman [1959]; all others are from this study. Triangle denotes NUVEL-1A plate vector calculated for Hokuri Creek [DeMets et al., 1994]. Notice the distinct dips and slip vectors of the southern and central Alpine Fault. (C) Field photos showing examples of fault rocks overthrusting Quaternary sediments and geomorphic expression of fault motion. Up/Down (U/D) denotes sense of dip-slip fault motion and paired arrows denote sense of strike-slip motion. PSS stands for principal slip surface. Geomorphic observations are everywhere compatible with fault plane slickenline data.
2.3.3 Protoliths

2.3.3.1 Greenland Group

The Greenland Group is a widespread and voluminous metasedimentary unit of the Buller terrane which comprises a portion of the AUS plate. The Greenland Group is the principal lithology outcropping west of the southern and central portions of Alpine Fault as far south as McKenzie Creek (Figure 2.1B). The unit consists predominantly of interbedded quartzose metasandstone and metamudstone, variably metamorphosed to quartz-muscovite-biotite schist and amphibolite-facies gneiss [e.g., Rattenbury et al., 2010]. Pelitic layers in the gneiss often exhibit coarsely-crystalline boudinaged quartz-feldspar-muscovite leucosomes. Local granite intrusions have created hornfels textures. In the vicinity of the Martyr River and Jackson Bay, there is a regional increase in metamorphic grade from chlorite zone to sillimanite-microcline zone to the southeast towards the Alpine Fault [e.g., Mortimer et al., 2012]. Within 160 m of the Alpine Fault at the Martyr River, the sillimanite grade gneiss is overprinted by a mylonitic fabric associated with chlorite alteration and an absence of biotite. Transposed leucosomes are still identifiable within the mylonitic fabric, and the age of this mylonitization is discussed in Section 2.6.

2.3.3.2 Brook Street Volcanic Group

The Brook Street Volcanic Group of the Brook Street terrane, occurs on the Pacific Plate in fault-bounded slivers adjacent to the Alpine Fault at the Martyr River and McKenzie Creek (Figure 2.1B). At McKenzie Creek, the unit is chloritically altered and dominated by andesite-derived metasandstone, metatuff, metaconglomerate and metabreccia [e.g., Rattenbury et al., 2010]. In places it has a well-developed foliation, while in others the unit can be chert-like and completely lack any obvious anisotropy. At the Martyr River, the Brook Street Volcanic Group has a mylonitic fabric. Here the rock consists solely of fine-grained chlorite-epidote with occasional quartz segregations. Epidote-rich horizons are boudinaged. It contains rare 1.5 m-wide by 6 m-long foliation-parallel pods of a mylonitized diorite with large hornblende porphyroclasts. The age of this mylonitization is discussed in Section 2.6.

2.3.3.3 Mt Webb Gneiss

The Mt Webb Gneiss is an enigmatic unit consisting of biotite garnet gneiss, mylonitic amphibolite, gneissic biotite granite and lenses of marble outcropping in the vicinity of Hokuri and McKenzie creeks (Figure 2.1B) [e.g., Turnbull, 2000]. It is characterized by low
regional abundance of faults and fractures (< 5 through-going fractures per 1 meter length of outcrop). Textural and metamorphic facies distinctions led Ballard [1989] to propose a new unit despite lithologic similarities to the Thurso Formation of the Anita Shear Zone [Wood, 1962]. However, based on our textural and lithologic observations, I concur with Wood [1962]’s original correlation. Although not exposed adjacent to the Alpine Fault at Hokuri Creek due to Quaternary sediment cover, the Mt Webb Gneiss is exposed in creeks to the southwest of the McKenzie Creek locality where it appears relatively unfractured and unaltered.

### 2.3.3.4 Kaipo Mélange

In this study, I identify, describe and formally name the Kaipo Mélange, an AUS plate-hosted tectonic mélange which crops out adjacent to the Alpine Fault for ~40 km from the Pyke River to Milford Sound (Figure 2.1B). Because this unit forms the immediate footwall at two key study sites and has a genetic relationship to fault core rocks, I briefly describe it here based on reconnaissance mapping done at six sites between the Kaipo Slips and McKenzie Creek.

Outcrops of this mélange had previously been examined by Wellman and Wilson [1964] who introduced the stratigraphic name “Kaipo Formation” and did not recognize it as a mélange. At their type section, the Kaipo Slips, the section includes sequences of granite between fault-bounded sedimentary units without any evidence for stratigraphic continuity or contact metamorphism. Here, I use the name “Kaipo Mélange” because I observe the sequence everywhere along strike to be strictly a tectonic mixture including exotic blocks [as per Raymond, 1984].

In southern exposures (e.g., Kaipo Slips, Alteration Creek), large (up to 70 m-across) fault-bounded blocks of chlorite-altered felsic intrusives, deformed Middle Eocene-Late Oligocene limestone [Nathan, 1978; Sutherland et al., 1996], calc-silicate, and rare serpentinite occur in a scaly calcareous mudstone matrix (Figure 2.3A, 2.3B, and 2.3C). Northeastern exposures (e.g., Hokuri Creek, McKenzie Creek) contain smaller (< ~5 m-across) tabular blocks of serpentinite (common), felsic intrusives, quartzose schist, marble, and limestone (rare) in a sheared phyllosilicate matrix (Figure 2.3D and 2.3F). Pervasive alteration of blocks after incorporation in the mélange makes a direct comparison to source rocks difficult; provenance of the felsic intrusives, schist and serpentinite remains unknown.
Figure 2.3  Kaipo Mélange. Field photographs contrasting Kaipo Mélange textures at the Kaipo Slips (A–C) in the southwest and Hokuri Creek (D–F) in the northeast. Key along-strike changes towards the southwest include: a decrease in block size, increase in block aspect ratio, less sedimentary and plutonic blocks, more serpentinite blocks, and a more planar phyllosilicate matrix (in contrast to folded, fissile, calcareous mudstone matrix to the southwest). Some blocks have been outlined for clarity. Abbreviations used: g (granite), m (calcareous mudstone or phyllosilicate mélange matrix), s (serpentinite), ls (limestone), q (quartz), d (diorite), p (pegmatite), as (altered schist).
From northeast to southwest, the mélange increases in width, reaching a maximum horizontal width of over 1.5 km. From northeast to southwest, average block size increases and block aspect ratio decreases. Southwestern exposures tend to have a higher percentage of incompetent matrix to competent blocks; in addition, the incompetent phyllosilicate matrix is more chaotically folded (cf. Figure 2.3C to 2.3D and 2.3F). Southwestern exposures also lack the boudinaged blocks, brittle fractures, and through going shear surfaces recorded in the northeastern outcrops (cf. Figure 2.3A, 2.3B, and 2.3C to 2.3D, 2.3E and 2.3F); these observations indicate that the deformation history of the mélange varies along strike. I note that Sutherland and Norris [1995] incorrectly attributed basement rocks exposed northwest of the fault core at Hokuri Creek as cataclyastically deformed upper greenschist or amphibolite grade Greenland Group; these footwall basement rocks are a part of the Kaipo Mélange (e.g., Figure 2.3D). Tectonic implications of this mélange are explored in Section 2.7.

2.3.4 Damage zone

On the southern Alpine Fault, steep northwest-draining creeks cut orthogonally to the Alpine Fault, and provide the best outcrops of both sides of the plate boundary structure through the entire width of the damage zone. These outcrops were documented with photographs, structural measurements and sampling. I find that the extent and nature of the damage zone varies significantly along strike.

2.3.4.1 Martyr River

At the Martyr River, the damage zone width is asymmetric: ~160 m-wide within the AUS plate and ~80 m-wide within the PAC (Figure 2.4). Foliations in the Greenland Group gneiss and mylonite both tend to strike parallel to the Alpine Fault, but dip 60–80° to both the northwest and southeast (Figure 2.4A, 2.4B, and 2.4C). The mylonitic fabric tends to dip more steeply than the gneiss and has a weak chlorite alteration overprinting it. Secondary faults containing phyllosilicate gouge can usually be constrained to have < 1 m of offset; most occur within about 40 m of the Alpine Fault core. Late-stage tensile fractures up to 1 m in length commonly form orthogonal to the mylonitic foliation (Figure 2.4B).

Within 3–10 m of the fault core, Greenland Group mylonites are overprinted by a foliated cataclasite exhibiting a strong S-C fabric defined by phyllosilicate-rich S shear planes parallel to the Alpine Fault, inclined phyllosilicate-rich C planes and fractured, sigmoidal clasts of the quartz-rich portions of the Greenland Group mylonite (Figure 2.4C). This S-C relationship indicates dextral-normal motion consistent with expected Alpine Fault kinematics. Calcite
Figure 2.4  Key features of the Alpine Fault damage zone at the Martyr River. Planar structures are represented as poles to planes on equal area lower hemisphere stereonets. The subscripts AUS and PAC are used to denote Australian and Pacific plate structures. Foliations with an S, planes of shear are denoted with a C, and tensile fractures are denoted with a F. Black bars above the schematic cross section denote areas where there is no outcrop. Greenland Group (GG) gneiss (A) is overprinted by a Cretaceous mylonitic fabric (B), which is then overprinted by minor faults and fractures of the Alpine Fault zone, including a strong S-C cataclastic fabric within 4 m of the fault core (C). Note transposed leucosomes are still identifiable within the mylonitic fabric (cf. B). The pre-existing mylonitic foliation tends to form the long axis of the clasts within the S-C cataclastic fabric, parallel to the S planes. Brook Street Volcanic Group (BSVG) rocks have an Early Miocene mylonitic fabric overprinted by a chlorite-epidote alteration zone, which is then overprinted by minor faults and fractures of the Alpine Fault zone (D), especially protocataclasite (E), and cataclasite (F) within ~30 cm of the principal slip zone (PSS). Discontinuous open fractures in the PAC plate (e.g., e) were not mapped in detail due to their short lengths, but are generally oriented perpendicular to the foliation. Stars denote locations of \( ^{40}\text{Ar}/^{39}\text{Ar} \) geochronology samples discussed in the text; see Section 6 for discussion of the ages of mylonitization.
veins only occur crosscutting the mylonite clasts, indicating that they pre-date the S-C fabric. Rare gouge zones cross-cut the S-C fabric at a high angle suggesting the S-C fabric is no longer active. A c. 2 ka iron-cemented fluvial gravel unit deposited on a portion of this outcrop does not show any signs of deformation. Above this iron cemented fluvial gravel unit, < 2 ka fluvial gravels have been passively overridden by ~10 m by the principal slip surface of the Alpine Fault (Figure 2.2C).

On the PAC plate, Brook Street Volcanic Group mylonites are openly folded on a cm-dm scale (Figure 2.4D). These mylonites have a pervasive chlorite alteration within 40 m of the fault core; this alteration frequently obscures the mylonitic fabric except where quartz-rich mylonitic segregations occur. Calcite, quartz and chlorite coat many of the fracture and fault surfaces. Though fractures and minor faults are abundant throughout the PAC plate alteration zone, remnant foliations are continuous for over 5 m, indicating that chemically altered damage zone fault rocks are best described as protocataclasites (Figure 2.4E). Cataclasite, which contains fragmented particles that have been translated and rotated, occurs only within ~1 m of the fault core. Foliated cataclasite, with a network of phyllosilicate-rich shear bands that form an S-C fabric consistent with dextral shear, occurs within ~30 cm of the fault core (Figure 2.4F).

On both sides of the fault at the Martyr River, I observe that fractures are most pervasive in the most altered rocks. In the vicinity of the Martyr River, a well-indurated, iron-oxide cemented glacial till (inferred to be c. 10 ka) overlies many fault-damaged outcrops; this till acts as a deformation marker that delimits the extent of fault damage at the surface (e.g., Figure 2.4D). Fractures and small offset faults in the bedrock nowhere affect the cemented till above the bedrock. This glacial till is not offset across foliated cataclasites adjacent to the fault core on the AUS plate, suggesting they are not actively deforming. The small faults and fractures crosscutting the PAC mylonites at a distance of 60 m from the fault core are similarly inactive. A < 18 ka glacial silt unit occurs on the PAC plate at the Martyr River within 20 m of the fault core; it exhibits folds resulting from soft sediment deformation which have been later cut by hairline reverse faults with cm-scale offset. The active damage zone at the surface at the Martyr River appears to be highly localized (potentially < 40 m-wide).
2.3.4.2 McKenzie Creek & Hokuri Creek

At McKenzie Creek and Hokuri Creek, the damage zone is considerably harder to define as it is unclear to what extent alteration and fracturing in the Kaipo Mélange can (and should) be directly attributed to the Alpine Fault. During reconnaissance mapping, detailed studies were mostly focused on the fault core. Nevertheless, the width of the zone appears to be much less than at the Martyr River and I define the damage zone as being ~90 m-wide (40 m-wide within the AUS plate and 50 m-wide within the PAC plate).

The PAC plate damage zone at McKenzie Creek is pervasively fractured. Here, tensile fractures break the isotropic chert-like Brook Street Volcanic Group metatuff into < 1–2 cm-sized cubes within 2 m of the PSZ (Figure 2.5B). These fracture sets are everywhere unmineralized. The PAC plate basement (Mt Webb Gneiss) is not exposed adjacent to the Alpine Fault at Hokuri Creek, but Mt Webb Gneiss is exposed adjacent to the fault at an unnamed tributary southwest of McKenzie Creek. Here Mt Webb Gneiss displays 2–5 cm-spaced tensile fractures that are orthogonal to the foliation, few faults, and little chloritic alteration. Chlorite alteration is strongest in the PAC plate Brook Street Volcanic Group lithologies within 20 m of the fault core at McKenzie Creek; overall, chloritic assemblages are less pervasive at McKenzie and Hokuri Creeks. Friction melt (pseudotachylyte) is not observed at any of the three sites in the study.

At both Hokuri and McKenzie creeks, secondary faults within 200 m of the fault core exhibit little evidence of brittle cataclasis and accompanying grain size reduction. These secondary faults typically have displacements ≤ 1 m and do not offset capping Quaternary sediments. Hokuri Creek, in particular, contains a record of Quaternary glacial, fluvial, and lacustrine sediments that have been studied in detail by Sutherland and Norris [1995] and Berryman et al. [2012a]. Sutherland and Norris [1995] describe flat-lying 15 ka glacial lake silts which coat AUS plate outcrop along the Hokuri River for a distance of about 60 m from the fault core; these silts are unfaulted, indicating the secondary faults exposed within the underlying Kaipo Mélange have been inactive over this time. The sequence of post c. 10 ka sequence of peat and silt on the PAC plate at Hokuri Creek are undisturbed and unfaulted, except where they were close enough to the fault scarp to have interfingering colluvial wedges [Berryman et al., 2012a]. It is worth noting that Sutherland and Norris [1995] describe glacial silts dipping 20° both on the AUS plate and PAC plate at Hokuri Creek; these glacial silts were folded into an anticline-syncline pair at some time during the last 100 kyr.
Because I do not observe secondary faults in the damage zone with significant offset or evidence of brittle grain size reduction, and because no secondary faults disturb capping Quaternary sediments (apart from those within the immediate hanging wall at the Martyr River), I argue that the vast majority of slip on the Alpine Fault at the surface must be accommodated within a single fault core or on its margins. At all three locations mapped in detail, I observe Quaternary sediments juxtaposed in fault contact against at least one margin of the fault core. Since the primary concern of this chapter is characterizing slip on the southern Alpine Fault, I now focus on a detailed examination of fault core exposures.

2.3.5 Fault core

I present the first complete composite cross sections through the fault core of the southern Alpine Fault from exposures at three locations (Figure 2.5A, 2.5B, and 2.5C) compiled from field measurements, measured sections and high-resolution orthorectified field photo tracings (e.g., Figure 2.6A, 2.6B, and 2.6C). At all three locations, the fault core is a single, well-defined, > 1 m wide tabular zone containing only AUS plate derived materials, indicating that the geologically defined sensu stricto plate boundary is the striated southeastern margin of the fault core.

At the Martyr River, a 1.5 m-wide fault core is composed primarily of blue-grey and white foliated chlorite-illite gouge (according to XRD analyses presented in Section 2.4) with < 4 cm-long clasts of quartz + calcite (Figure 2.6A). Micro-folded mylonitic textures with cross-cutting calcite present in many of the clasts with diameters larger than 3 mm are identical to those of quartzopelitic mylonites on the adjacent AUS plate. Cataclastically deformed sigmoidally shaped clasts indicate a dextral-normal shear sense (Figure 2.6D). Chlorite-illite gouge foliation is sub-parallel to the fault core margins.

At the southeast margin of the fault core, AUS plate quartz-pelite-derived ultracataclasite, sheared rock and phyllosilicate gouge are juxtaposed against altered chloritic PAC plate cataclasite; this principal slip surface coincides with a prominent surface trace of the fault and is inferred to be the active PSS. Intense surface weathering precluded collection of this material for further analysis. On the northwest margin of the fault core, I mapped a 15 cm-thick blue illite-chlorite gouge with fine 2 mm-wide laminations and an average quartz + calcite clast size of < 1 mm; this gouge is interpreted to represent a former PSS and was collected for friction and permeability experiments (described in Section 2.5). Few other outcrops of this quartz-pelite-derived gouge were observed, but its Greenland Group parent
lithology is widespread along the fault to the northeast and southwest, and it may be extensive along strike.

Between McKenzie Creek and Hokuri Creek, deformation is strongly localized in a 1.1 to 12 m-wide fault core containing pods of completely serpentinized peridotites (serpentinites) suspended in trioctahedral smectite (saponite)-rich foliated fault gouge. Surface outcrops of these serpentinite-derived gouges can be traced for 10 km along strike. Uncommon
cataclased sigmoidal clasts consistently indicate dextral movement (Figure 2.6E). Clast size, clast aspect ratio and fault core thickness consistently increase to the southwest. The largest clasts or pods (sometimes > 5 m across) tend to occur in the center of the fault core and commonly are surrounded by a reaction halo of lizardite-chlorite gouge (Figure 2.6C and 2.6F). These pods typically have oblate shapes (e.g., dimensions of 2.5 m x 2.5 m x 1.5 m) with their long axes parallel to the fault core margins. While clast concentrations vary throughout the fault core, the regions surrounding large serpentinite pods tend to be more clast-rich. I did not observe saponite gouge outside the fault core. Outcrops where the fault core is less than 2 m-wide exhibit gouge foliations concordant to the fault core margins, but foliations within wider cores are discordant to the margins (Figure 2.5). Saponite gouge foliation anastomoses around the rigid serpentinite clasts. Within the gouge there are no obvious through-going zones or planes on which slip may have been localized.

2.4 Mineralogical and Microstructural Observations

2.4.1 Analytical methods

Bulk rock quantitative x-ray diffraction (XRD) was used to determine the matrix mineralogy of each fault core gouge (see Boulton et al., 2012 for detailed analytical methods; analyses by Mark Raven at CSIRO). Petrographic thin sections of fault gouges were prepared dry and polished using 1 µm diamond paste. Thin sections were imaged in a Zeiss Sigma field emission scanning electron microscope (SEM), mostly using an angle selected backscattered (AsB) detector to highlight compositional and topographical variations. Energy-dispersive X-ray (EDS) was used to identify minerals (paired with optical microscopy and XRD data) and create element variation maps. The EDS was particularly useful for distinguishing between minerals that could not be isolated by XRD, like muscovite and illite. Thin section petrography aided determination of the variety of serpentine present (lizardite).

Figure 2.6 Fault Core Observations. Field photographs of the Alpine Fault core at (A) Martyr, (B) McKenzie and (C) Hokuri sites (hammer, hammer, and backpack for scale respectively). PSS stands for principal slip surface. (D, E, F) Hand-sample-scale photos of clast textures suspended in phyllosilicate-rich gouge matrix. Srp stand for serpentinite and Qtz stands for quartz. (G, H, I) Petrographic thin section photomicrographs of phyllosilicate-rich fault gouges seen in cross polarized light. Dark areas are where phyllosilicate matrix is missing due to sample preparation. (J, K, L) Scanning electron microscope (SEM) images of phyllosilicate-rich fault gouges taken with an angle selected backscattered (AsB) detector. Black voids are desiccation cracks due to sample preparation.
2.4.2 Martyr River

The bulk of the exposed AUS Plate fault core at the Martyr River is composed of macroscopically-foliated gouge with a chlorite-illite matrix. The softest gouge in the fault core, most readily malleable by fingertips, is interpreted to be a former PSS or perhaps secondary PSS; this gouge is illite-chlorite-rich, with illite being more abundant than chlorite (Table 2.2). Microstructurally, the main textural difference between the foliated chlorite-illite and illite-chlorite gouges is that clasts are relatively abundant within the stronger foliated gouge; these clasts inhibit the formation of interconnected phyllosilicate layers. Although some common extinction (and therefore preferred orientation) of phyllosilicate grains is indicated in photomicrographs (Figure 2.6G), SEM images (Figure 2.6J) reveal that that individual phyllosilicate grains are seldom particularly continuous. Instead, the phyllosilicates form packages a few microns across, bounded by discrete, anastomosing surfaces. Phyllosilicates within individual packages have common orientations, but these orientations are not parallel in adjacent packages.

Both Martyr River fault core gouges contain clasts of polycrystalline quartz ± illite ± albite ± cross-cutting calcite, and monomineralic clasts of titanite and rare pyrite (Figure 2.7A). This is the same assemblage found in the AUS plate quartzopelitic mylonites immediately northwest of the fault core (with the exception that in the AUS plate mylonites muscovite occurs in place of illite). Some clasts have lobate (i.e., embayed) boundaries. Mylonitic textures are visible in some of the larger clasts, and cross-cutting calcite veins within the clasts are truncated by the phyllosilicate foliation at clast boundaries. These veins are similar to those pervasively present in the adjacent AUS plate quartzopelitic rocks. No calcite veins or disseminated grains are observed within the gouge matrix. Clasts 1–2 cm across commonly have sigmoidal tails of comminuted clast material, while clasts smaller than 1 mm tend to be more equant with sub-angular to sub-rounded shapes (Figure 2.6D, 2.6G, and 2.6J).

2.4.3 McKenzie Creek and Hokuri Creek

McKenzie Creek and Hokuri Creek gouges have a saponite-lizardite-chlorite matrix (Figure 2.7B and 2.7C). Both gouges contain over 60% trioctahedral smectite (saponite). McKenzie Creek and Hokuri Creek gouges are petrographically similar, but differ in minor proportions of saponite, chlorite, quartz and talc (Table 2.2). In thin section, individual phyllosilicate layers are commonly traceable for several millimeters (Figure 2.6H and 2.6I). Gouge foliation can be planar or tightly folded. Individual phyllosilicate sheets are ~5 µm-long. Unlike the
Table 2.2  Bulk Rock Quantitative XRD

<table>
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<tr>
<th>CSIRO ID</th>
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<th>Quartz</th>
<th>Orthoclase/Microcline</th>
<th>Albite</th>
<th>Calcite</th>
<th>Smectite*</th>
<th>Mica/Illite</th>
<th>Chlorite</th>
<th>Serpentine (Lizardite)</th>
<th>Talc</th>
<th>Amphibole/Actinolite</th>
<th>Pyrite</th>
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* the smectite in samples 34111 and 34959 are trioctahedral and likely saponite

Martyr gouges, at the SEM scale, I do not observe an interconnected network of discrete surfaces (Figure 2.6K and 2.6L). The EDS mapping down to a few micron spot size show no detectable variations in Mg (or other elements) across foliation (Figure 2.7B and 2.7C), although I observe varying grayscale intensities indicating either compositional or topographic variations in extreme close-up AsB images (e.g., Figure 2.6L).

Clasts are typically ellipsoidal in shape with the long axis parallel to the gouge foliation and have sub-rounded to rounded surfaces (Figure 2.6K and 2.6L; Figure 2.7B and 2.7C). Serpentinite clasts occasionally preserve original peridotite texture, but typically these are obscured by serpentinization textures (Figure 2.6I). I observe scale dependence in clast mineralogy in the Hokuri and McKenzie Creek fault cores. Outcrop to thin section scale clasts are dominated by serpentine minerals, with rare quartzopelitic and quartzfeldspatic lithogies present. In contrast, at the SEM scale (about < 50 µm) calcite and quartz are the most abundant clast phases (although not volumetrically dominant). Many of the larger serpentinite pods seen in outcrop are surrounded by a < 10 cm-wide reaction rind dominated by lizardite with associated chlorite (Table 2.2; Figure 2.6C and 2.6F). In thin section, a highly birefringent rim similar to the macroscale rind can be found around serpentinite clasts; rim-forming minerals have been dragged into tails consistent with a dextral shear sense (Figure 2.6H and 2.6I). Lizardite-rich gouge clasts with a diameter less than 1 mm have sometimes been incorporated into the saponite-rich gouge, but never vice versa.
Figure 2.7  Fault Core Mineralogy. Whole rock quantitative x-ray diffraction (XRD) data, energy-dispersive X-ray spectroscopy (EDS) element maps (least-squares averaged), and angle selected backscattered (AsB) SEM images of phyllosilicate-rich gouges from (A) Martyr, (B) McKenzie, and (C) Hokuri creeks. The field of view for all three images of a sample is the same. Black voids are desiccation cracks due to sample preparation. Mineral abbreviations used: Ab (albite), Act (actinolite), Ap (apatite), Cal (calcite), Fsp (feldspar), Ill (illite), Lz (lizardite), Mag (magnetite), Py (pyrite), Qtz (quartz), Sap (saponite), Tlc (talc), Ttn (titanite).
The clast population and mineral assemblage is diverse, reflecting the diversity of lithologies present in the protolith Kaipo Mélange. Clast mineralogies include: serpentinite ± actinolite ± talc ± quartz ± pyrite ± magnetite ± calcite; calcite; quartz ± calcite; titanite; epidote; chlorite; magnetite; pyrite; chromite ± magnetite; chalcopyrite; and rare apatite + titanite + talc + chlorite + saponite; and quartz-feldspar-muscovite-biotite schist lithics. I observe clasts of chromite with rims and internal cracks of magnetite, similar to those observed by Moore and Rymer [2012] in saponite-rich gouge from SAFOD and a nearby surface outcrop. The presence of remnant apatite and titanite in some clasts associated with talc and saponite suggests the clasts are former mafic rocks that have been severely metasomatically altered; these otherwise low-Mg clasts are the only clasts I have identified saponite within (Figure 2.7C).

2.5 Frictional and Hydrological Properties

2.5.1 Experimental methods

Friction and permeability experiments were undertaken on intact wafers of fault core gouges. Wafers had final dimensions of 5–8 mm thick, 54 mm wide and 61 mm long. They were cut with the long-axis parallel to fault shear direction. In each experiment, two fault rock wafers were placed between a three-piece steel block assembly, jacketed, and deformed at room temperature in a servo-controlled biaxial testing apparatus fitted with pressure vessel (see Samuelson et al. [2009] for deformation apparatus and experimental method details) (Figure 2.8 inset). Four friction experiments were conducted on fault gouges collected from the Martyr River, McKenzie Creek and Hokuri Creek sites. An additional experiment was done on fault gouge collected from a fresh exposure at Gaunt Creek on the central section of the Alpine Fault, to allow a more direct comparison between fault gouges exposed on the central and southern Alpine Fault (Table 2.3).

At the start of each experiment, wafers were saturated with pore fluid (University Park, Pennsylvania, USA tapwater), loaded to a constant effective normal stress between 6 MPa and 31 MPa and left under load until compacted thickness remained unchanged. Wafers were then sheared by driving the center block at a constant displacement rate to attain steady state frictional behavior. The steady state coefficient of friction ($\mu_{ss}$) was calculated as the ratio of shear stress ($\tau$) to effective normal stress ($\sigma_n'$) assuming zero cohesion.
In the second part of each experiment, velocity step tests were conducted by varying the load point velocity between 1 and 300 µm s⁻¹; velocity was increased incrementally and the frictional response to each velocity step was recorded. Empirical rate-and-state friction (RSF) equations were used to describe velocity step results, where

\[ \mu = \mu_0 + a \ln(v/v_0) + b \ln(v_0 \theta d_c) \]  

(1)

and \( \mu_0 \) is the friction coefficient at some reference sliding velocity \( v_0 \), \( v \) is the sliding velocity, \( a \) is a parameter that describes the direct effect, \( b \) is the parameter that describes the evolution effect, \( d_c \) is the critical slip distance, and \( \theta \) is the state parameter, given by

\[ d\theta/dt = 1 - v \theta d_c \]  

(2)

in the Dieterich formulation of the law [Dieterich, 1979]. The friction rate parameter \((a-b)\) describes the effect of changes in slip velocity on steady state friction, and from (1) and (2), \((a-b) = \Delta \mu/\Delta \ln v\). The friction rate parameter for each velocity step was determined by iteratively solving equations (1) and (2) to find least squares best fit to the data [e.g., Marone, 1998].

At the end of each experiment, a series of slide-hold-slide tests were done to measure the frictional healing. Frictional healing \((\Delta \mu)\) is the difference in peak friction following a hold relative to the steady-state sliding friction prior to the hold. In these tests, wafers were sheared at 10 µm s⁻¹, held motionless for a prescribed time \((t_h)\) between 1 and 1000s and then driven at 10 µm s⁻¹ again.

Separate flow-through permeability measurements were made on wafers of each fault rock at 31 MPa effective normal stress using the same experimental configuration. In the pressure vessel, normal and confining pressures were applied and maintained via hydraulic servo-controllers. Sample wafers were then saturated with pore fluid and allowed to equilibrate until steady state boundary conditions were reached. A differential fluid pressure was imposed to induce flow normal to the shear direction of the intact wafers, and permeability was calculated at steady state flow rate using Darcy’s law.
Table 2.3  Friction / Permeability Experiment Details* and Results

<table>
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<tr>
<th>Experiment</th>
<th>Sample</th>
<th>$\sigma_n'$ (MPa)</th>
<th>$P_c$ (MPa)</th>
<th>$P_p$ (MPa)</th>
<th>$k$ (m$^2$)</th>
<th>$\mu_s$</th>
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<td>p3673</td>
<td>Gaunt Creek gouge</td>
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<td>1</td>
<td>3.10 e-20</td>
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<td>p2861</td>
<td>Martyr River gouge</td>
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<td>p3152</td>
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</table>

* Effective normal stress ($\sigma_n'$) represents the combined effects of applied normal load, confining pressure ($P_c$) and pore pressure ($P_p$).

2.5.2  Frictional properties

All fault core rocks tested have low friction coefficients ($\mu_{ss} = 0.12$ to 0.37) and positive values of $a-b$ between 0.0035 and 0.015 (see Figure 2.5 for locations of samples; Table 2.3; Figure 2.8A). The weakest fault gouges, collected from McKenzie and Hokuri Creeks, are composed primarily of saponite, chlorite and lizardite (Figure 2.8; Table 2.2). The very low friction coefficients of these gouges ($\mu_{ss} = 0.13$ and 0.12 respectively) lie within the range measured on saturated saponite-rich gouges collected from the Gokasho-Arashima Tectonic Line, Japan ($\mu_{ss} = 0.06 – 0.12$) [Sone et al., 2012] and the San Andreas Fault ($\mu_{ss} = 0.09 – 0.25$) [Lockner et al., 2011; Carpenter et al., 2011, 2012]. Saponite-rich fault core rocks also exhibit velocity-strengthening behavior, with positive values of the friction rate parameter ($a-b = 0.0049 – 0.0098$) that fall within the range reported for fault gouge comprising the actively creeping central deforming zone (CDZ) of the San Andreas Fault ($a-b = 0.0040 – 0.019$) [Carpenter et al., 2012].

Phyllosilicate-rich muscovite/illite-chlorite fault gouge from Gaunt Creek and saponite-chlorite fault gouges from McKenzie and Hokuri Creeks exhibited frictional healing of zero or below, even over the longest hold times ($t_h = 1000$ s) (Figure 2.8B). The healing behavior of illite-chlorite gouge and chlorite-illite foliated gouge from the Martyr River varied. The illite-chlorite gouge consistently strengthened with hold time, while the foliated gouge weakened over short hold times ($t_h = 3, 10, and 30$ s) and strengthened over longer hold times ($t_h = 300$ and 1000 s). A lack of frictional healing with hold time was observed in muscovite/illite-chlorite gouges on the central section of the Alpine Fault [Boulton et al., 2012],
Figure 2.8  Fault Core Friction Parameters. (A) Plot of friction rate parameter \((a-b)\) versus upstep velocity \((\mu m/s)\) for intact wafers of phyllosilicate-rich fault gouges. See text for definitions and Figure 2.7 caption for mineral abbreviations. Experiment names: Gaunt Creek muscovite/illite-chlorite gouge \((p3673)\), Martyr River illite-chlorite basal gouge \((p2861)\), Martyr River chlorite-illite foliated gouge \((p3152)\), McKenzie Creek saponite gouge \((MK\_CGp3373)\), and Hokuri Creek saponite gouge \((p3372)\). (B) Plot of frictional healing \((\Delta \mu)\) versus hold time \((s)\). Inset shows double direct shear apparatus configuration used for these experiments.

and saponite-chlorite-rich fault gouges on the San Andreas Fault [Carpenter et al., 2011, 2012]. Fault gouges that do not undergo frictional healing are requisite for aseismic creep, but there currently exists no detailed microphysical understanding of this behavior [e.g., Scholz, 2002].

2.5.3 Fault gouge hydrological properties

Fault core rocks had very low fault perpendicular permeability (Table 2.3). The saponite-rich fault core gouges from McKenzie and Hokuri Creeks were less permeable \((k = 1.5 \times 10^{-22}\)
and \( k = 3.6 \times 10^{-21} \), respectively) than the illite-rich gouge (\( k = 1.88 \times 10^{-20} \)) and chlorite-illite foliated gouge (\( k = 1.16 \times 10^{-19} \)) collected from the Martyr River. SEM observations show greater interconnectedness of phyllosilicates in the McKenzie and Hokuri Creek gouges than in Martyr River gouges, which is consistent with the lower permeability of these gouges. These permeability values are of the same order of magnitude as or lower than those published on central Alpine Fault gouges [Boulton et al., 2012], San Andreas Fault gouges [Morrow et al., 2011], Median Tectonic Line gouges [Wibberley and Shimamoto, 2003], and Nojima Fault Zone gouges [Lockner et al., 2000]. Low fault core permeability may weaken the Alpine Fault interseismically and coseismically because it allows for the pressurization of pore fluids [Lachenbruch, 1980; Sleep and Blandpied, 1992; Faulkner and Rutter, 2001; Rice, 2006; Garagash, 2012].

2.6 \(^{40}\text{Ar}/^{39}\text{Ar}\) Age Constraints on Fault Zone History

\(^{40}\text{Ar}/^{39}\text{Ar}\) step heating analysis was undertaken on individual muscovite crystals from four Greenland Group (AUS plate) samples from the Martyr river area (Table 2.4; Figure 2.4). Directly equivalent analyses could not be obtained from the PAC plate side of the Alpine Fault in this area due to the absence of appropriate mica phases, but comparable data were derived from step heating of a 30 mg multi-grain aliquot of hornblende separated from a sheared diorite hosted within Brook Street Volcanic Group (PAC plate) mylonites. More detailed analytical methods, petrology and results are provided in Supplement 2.

Within the AUS plate domain, all four mica age spectra express a common general form. Young initial ages increase to older ages over the bulk of gas release, defining an irregular plateau or exhibiting mild dispersal about a general attractor (Figure 2.9A). These results are consistent with the deformed character of the sampled materials. Deformed materials are typically partially disturbed systems with a thermal and/or mineralogical overprint that resets the argon systematics of exterior and less retentive sites within the crystal lattice [Batt et al., 2004; Beltrando et al., 2009; Forster and Lister, 2009]. In such systems, the young initial ages in particular can provide significant geological meaning by dating the nominal episode of disturbance. Although rendered imprecise by diffusional mobility of \(^{40}\text{Ar}\) during the disturbance episode, the older age attractor can also provide a qualitative guide to the inherited age upon which the disturbance was imposed [Beltrando et al., 2009; McDougall and Harrison, 1999].
Table 2.4  Summary of Analyzed Samples

<table>
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<tr>
<th>Sample Name</th>
<th>Analysis</th>
<th>Locality</th>
<th>Unit</th>
<th>Unit type</th>
<th>Latitude (°S)</th>
<th>Longitude (°E)</th>
<th>Elevation (m)</th>
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</tbody>
</table>

The AUS plate sample fabrics identified and dated in this study are all older than 70 Ma. These results indicate that mylonites on the AUS plate adjacent to the southern Alpine Fault appear to have been unaffected by Alpine Fault-related mylonitization beginning c. 25 Ma [Kamp, 1986; Cooper et al., 1987; Sutherland et al., 2000]. Characterization of the isotope systematics using inverse isochron plots could not be applied to the four AUS plate muscovite samples, as data from these tend to cluster near the radiogenic axis and exhibit a poor spread along nominal mixing lines, preventing the robust distinction of isotopic trends.

An Alpine Fault related signal is apparent, however, in the age signature of PAC plate amphibole sample OU82946. The inverse isochron plot \( ^{39}\text{Ar}/^{40}\text{Ar} \) ratio against \( ^{36}\text{Ar}/^{40}\text{Ar} \) ratio in Figure 2.9B characterizes the isotope systematics and provides additional confidence in these ages. Fitting reverse isochrons through the hornblende data produces refined model ages of 18 ± 4 Ma for the initial plateau of the PAC plate mylonite, which I infer results from resetting during Alpine Fault-related mylonitization under greenschist facies conditions (Figure 2.9B). The older 294 ± 10 Ma age obtained from this sample is interpreted as a Permian crystallization age for the hornblende diorite. Thus, the PAC plate amphibole sample records a Miocene disturbance age and an inherited Permian age.
2.7 Discussion

2.7.1 Timing of slip

2.7.1.1 Martyr River

Thermochronological constraints from $^{40}\text{Ar}/^{39}\text{Ar}$ dating of muscovite separated from AUS mylonites on the southern Alpine Fault reveal that ductile mylonitic fabrics identified in AUS plate rocks west of the Alpine Fault at Martyr Creek are older than 70 Ma. These results corroborate previous geophysical, geological, and geochronological data that indicate that the Alpine Fault exploits, at least locally, a pre-existing continental fault zone active during the Cretaceous (continental because it affects Greenland Group continental sediments) [e.g., Batt et al., 2004; Rattenbury, 1987; Sutherland et al., 2000]. Although younger AUS plate mylonites that post-date the inception of the modern Alpine Fault at c. 25 Ma [Sutherland et al., 2000] may be present at depth, my results demonstrate that Miocene-Recent exhumation has not been sufficient to expose such rocks along the southern Alpine Fault.

Figure 2.9 $^{40}\text{Ar}/^{39}\text{Ar}$ Age Plots. (A) Age spectra plotting variation in the nominal age (Ma) for each heating step against cumulative release of $^{39}\text{Ar}$ (%) as a proxy for experimental progress towards complete outgassing. (B) Inverse isochron plot ($^{39}\text{Ar}/^{40}\text{Ar}$ ratio against $^{36}\text{Ar}/^{40}\text{Ar}$ ratio) for sample OU82946 (PAC plate mylonite).
In contrast, hornblende $^{40}$Ar/$^{39}$Ar ages from mylonitized PAC plate diorite sample OU82946 collected 1 km northeast of the Martyr River indicate the ductile fabrics at this site were imposed at 18 ± 4 Ma – within the known period of Alpine Fault activity. These PAC mylonites were most likely a product of dextral-reverse shear on the central Alpine Fault. The reverse component of shear further north exhumed them at that location, and they were then dextrally translated to their present position in South Westland, where dominantly strike-slip fault motion could not accomplish this exhumation. PAC plate mylonites of comparable age on the central Alpine Fault have been removed due to high exhumation and erosion rates [e.g., Batt et al., 2004; Little et al., 2005]. PAC plate mylonites from near the Martyr River are thus some of the oldest exhumed mylonites with fabrics formed during activity on the Miocene-Recent Alpine Fault. The alteration and brittle deformation of the PAC plate mylonites at the Martyr River must also be related to Miocene-Recent Alpine Fault activity.

2.7.1.2 McKenzie Creek & Hokuri Creek

I only observe serpentinite pods hosted within Mg-rich Alpine Fault gouges in the vicinity of McKenzie and Hokuri Creeks, where the immediately adjacent AUS plate Kaipo Mélange contains the greatest abundance of serpentinized ultramafic lithologies. This observation, together with the textural similarities between fault-hosted and mélange-hosted serpentinites, indicates that the Kaipo Mélange is the source of the fault core lithologies. Thus, the serpentine-bearing fault core gouges must be younger than the mélange. At the Kaipo Slips, fold geometry and extent of re-crystallization in the Middle Eocene-Late Oligocene limestone blocks in the mélange is comparable to their counterparts exposed in stratigraphic succession 5 km to the northwest, indicating the Kaipo Mélange formed after limestone re-crystallization.

Nathan [1978] suggested that the thin-bedded mudstone and calcareous sandstone abundant at the Kaipo Slips is similar to Middle Miocene units exposed in a nearby coastal section (Kaipo Member of the Tititira Formation), which could place a further constraint on the maximum age of the Kaipo Mélange. I suggest that the mélange may represent subduction-related accretionary wedge material translated into the Alpine Fault from the southern extent of AUS plate continental crust. If the mélange formation was not a time-transgressive event along strike, the block provenance indicates that slip was localized into the current fault core here after the Late Oligocene (c. 24 Ma) and possibly after the Middle Miocene (13–8 Ma). A young age for the mélange is consistent with the youthful nature of the Alpine Fault offshore, which is characterized by fault segmentations and step-overs, changes in fault strike, and pull-
apart basins. These structural features appear to have formed as a result of inherited Eocene rift structures entering the subducting AUS plate, probably since 6–3 Ma [Barnes et al., 2005]. Sutherland et al. [2000] proposed that the Eocene rift boundary may be inherited from Cretaceous oceanic transform faults and an additional older discontinuity within continental Zealandia. The AUS mylonite data suggest this exploited discontinuity may be a crustal-scale, Cretaceous-active continental fault zone.

### 2.7.2 Nature of slip

#### 2.7.2.1 Saponite-rich gouges

Previous studies on the San Andreas Fault correlated creeping sections of the fault in central and northern California with serpentine-bearing rocks of the Franciscan Mélange [Allen, 1968; Hanna et al., 1972; Moore and Rymer, 2007], and the San Andreas Fault Observatory at Depth (SAFOD) drilling program demonstrated a direct correlation between active fault creep in two active fault cores at ~3200 m depth (the CDZ and SDZ) and saponite-bearing fault gouges [Bradbury et al., 2011; Carpenter et al., 2011, 2012; Holdsworth et al., 2011; Lockner et al., 2011]. The trioctahedral smectite saponite is thought to form by metasomatic reactions between quartzofeldspathic sedimentary rocks and serpentinite [e.g., Moore and Rymer, 2012 and references therein].

Within the Hokuri Creek fault gouge, clasts with relict apatite and titanite associated with talc and saponite may formerly have been mafic igneous rocks; these otherwise low-Mg clasts are the only clasts I have identified saponite within. The relatively low abundance of serpentinite clasts smaller than ~1 mm in the Mg-rich gouges may indicate these clast sizes preferentially react to form the saponite matrix. This would suggest that the saponite-forming reactions occur within the gouge zone itself. The thickness of the saponite-rich fault core at Hokuri Creek and McKenzie Creek, and the presence of lizardite-chlorite reaction rims surrounding some serpentinite pods, support a focused fluid flow origin for the saponite, as Holdsworth et al. [2011] proposed for the saponite-bearing SAFOD gouges.

#### 2.7.2.2 Mechanisms of slip

In all the examined gouges, the dominant through-going fabrics are associated with phyllosilicate minerals. Calcite veins are preserved only within clasts. This is dissimilar to observations in SAFOD fault gouge, which have been interpreted to indicate deformation accommodation by solution-transfer processes [Hadizadeh et al., 2012]. Furthermore, there is
no evidence of clast indentation, and only a few embayed clast boundaries. The latter could have formed by dissolution during contact with other clasts, but these contacting grains have since been translated away by shear within the surrounding matrix. Based on these observations, I infer that solution-transfer was not important during the most recent deformation of these materials.

The illite-chlorite gouges from the Martyr River are composed of spaced phyllosilicate grains between an anastomosing network of discrete surfaces. On the other hand, the saponite-lizardite-chlorite gouges are composed of continuous phyllosilicates sheets. I infer that the most recent increment of deformation within the illite-chlorite gouges occurred on the anastomosing surfaces and that the phyllosilicate packages were mostly passively rotated between these. Conversely, I infer that a more distributed deformation was relatively evenly accommodated by glide on weak phyllosilicate basal planes in the saponite-rich gouges. This is consistent with observations of experimental and natural serpentine gouge textures, where gouges produced by stable fault creep exhibited a strong alignment of phyllosilicates and lack of strain localization structures [Reinen, 2000]. In addition to mineralogy, I find that clast abundance and the presence of discrete surfaces likely exerts an important control on frictional properties, with the more clast-rich foliated gouge at the Martyr River exhibiting a higher friction coefficient and positive frictional healing over longer hold times.

On the central Alpine Fault, coseismic slip is inferred to have been focused within cm-thick layers of ultracomminuted material (e.g., ultracataclasite) forming the PSS at the footwall-hanging wall contact [Boulton et al., 2012; Sutherland et al., 2012]. I infer similar behavior at the Martyr River where cm-thick layers of ultracataclasite separate PAC and AUS plate lithologies; the ultracataclasite is frequently cemented, exhibits a single striated surface, and in places is juxtaposed against Quaternary sediments by fault motion. At the Martyr River, I can also clearly correlate surface traces of the fault with outcrops of the ultracomminuted material separating PAC and AUS plate lithologies.

In contrast, there are no planar layers of ultracomminuted material within the wide saponite-rich fault core at McKenzie Creek or Hokuri Creek. I consider it unlikely that a single event coseismic displacement of ~8 m could be accommodated within the fault core gouge without leaving evidence of slip localization. Therefore, I suggest that coseismic slip propagates along the slickenlined fault core margins at the contact between phyllosilicate-rich gouge and more competent lithologies. Ikari and Kof [2011] noted that fault core boundary interfaces have less cohesion that phyllosilicate-rich gouges, and results from field observations, laboratory
experiments, and numerical models all indicate that strain localization mostly commonly occurs between materials with different mechanical properties (competencies) [e.g., Goodwin and Tikoff, 2002 and references therein].

2.7.2.3 Extent of weak gouges

The extent to which the weak fault gouges accommodate displacement via fault creep and affect rupture propagation depends on the extent to which they are present along-strike and down-dip. My geological mapping indicates that exceptionally weak saponite-rich gouges derived from the Kaipo Mélange are present for at least 10 km along strike north of Lake McKerrow, but the mélange itself outcrops adjacent to the Alpine Fault for ~40 km onshore and may extend an additional 30 km offshore to Sutherland Sound where the continental shelf terminates against the Alpine Fault. Along-strike observations of the mélange indicate that the greatest abundance of ultramafic constituents occurs in the northeast where the tectonic fabric is more ductile in nature and marble is present instead of Tertiary limestone. I thus interpret the northeastern portion of the mélange to have been exhumed from a deeper structural level, and propose that the along-fault distribution of ultramafics (and thus saponite-rich gouges) may be greater than exposed at the surface. I have no direct constraints on the down dip extent of these gouges, except to note that pure smectite is unstable at temperatures above ~100°C, and is therefore unlikely to be present at depths greater than ~4 km assuming a normal geothermal gradient of 25°C/km [Huang et al., 1993]. Below this depth, the distribution of illite/muscovite, chlorite, mixed-layer chlorite-smectite, and serpentinite minerals, with correspondingly higher coefficients of friction, likely controls fault core strength [e.g., Moore and Lockner, 2011; Moore et al., 1996; Reinen et al., 1991, 1994; Schleicher et al., 2012].

2.7.2.4 Linking Fault Outcrops to Fault Behavior

The results of this study highlight a conundrum: weak fault rocks with textures and frictional properties suggestive of slow aseismic slip are exposed in surface outcrops of the Alpine Fault, despite abundant paleoseismic evidence that large magnitude earthquakes occur here [Cooper and Norris, 1990; Wells et al., 1999; Wells and Goff, 2007; Berryman et al., 2012a]. Although GPS, InSAR and survey data quality remain poor on the southern Alpine Fault, I have found no direct evidence of active fault creep in our study area, supporting previous observations made on the Alpine Fault overall by Evison [1971], Wood and Blick [1986], Wallace et al. [2007], Sutherland et al. [2007] and others. Phyllosilicate-rich fault core rocks examined in this study
are frictionally weak ($\mu_{ss} < 0.37$) and velocity strengthening ($a-b > 0$). In addition, all fault gouges exhibited healing rates of zero or below, reflecting static values of friction at or below sliding values of friction. Theoretically, velocity weakening frictional behavior and positive healing rates are required for earthquake rupture nucleation and propagation [e.g., Marone, 1998; Scholz, 2002]. Combined with previously published results, the experiments indicate that wherever studied, fault core gouges exposed on the surface on the Alpine Fault should creep aseismically [e.g., Boulton et al., 2012].

Adjacent to the locus of the widest (12 m-wide), frictionally weakest ($\mu_{ss} = 0.12$) saponite-bearing fault core at Hokuri Creek, an 8000 yr off-fault paleoseismological record suggests remarkably time-dependent rupture behavior involving large magnitude earthquakes [Berryman et al., 2012a]. Thus the fault core observations are in direct conflict with the known seismogenic behavior of the fault at the same location, which leads me to caution workers attempting to characterize the overall behavior of a fault based on surface outcrops. Where weak and strong materials are present in varying distributions [e.g., Handy et al., 2006; Perfettini and Ampuero, 2008; Fagereng and Sibson, 2010], fault roughness causes variations in normal stress [e.g., Dieterich and Smith, 2009], fluctuations in pore fluid pressure occur [e.g., Sibson, 1992; Hillers and Miller, 2007], or temperature, strain rate and grain-size sensitive deformation mechanisms compete with frictional sliding within the fault core [e.g., Sibson, 1977; Rutter, 1983; Bos et al., 2000, Bos and Spiers, 2002; Gratier et al., 2009, 2011], the Alpine Fault may display rheological behavior that varies spatially and temporally.

In particular, pressure- and temperature-dependent variations in fault rock rheology probably play an important role in the observed lack of aseismic creep. Experimental studies on a variety of fault gouge and fault gouge analogues reveal that temperature strongly influences a given material’s friction rate parameters, that is, whether it is velocity strengthening or velocity weakening in response to an instantaneous change in sliding velocity. Using different experimental apparatuses, confining pressures, pore fluid pressures, and sliding velocities between 0.01 and 100 $\mu$m/s, each of the following studies found that at elevated temperatures, velocity-strengthening gouges become velocity-weakening: wet quartz gouges (100°C to 300°C) [Chester and Higgs, 1992; Kanagawa et al., 2000], wet quartz-illite gouges (250°C to 400°C) [den Hartog and Spiers, 2012; den Hartog et al., 2012], wet granite gouges (100°C to 350°C, except at 250°C) [Blanford et al., 1995], and wet San Andreas Fault gouges (266°C to 349°C) [Tembe et al., 2009]. Above the critical temperature necessary for the activation of intracrystalline plastic deformation, $\sim$300±50°C [Tullis and Yund, 1977], quartz-
rich gouges in turn becomes velocity strengthening, a process thought to be responsible for the down-dip seismogenic limit [e.g., Sibson, 1982; Scholz, 2002]. Given the high shallow geothermal gradient of 60–70°C/km on the central Alpine Fault [Allis and Shi, 1995; Sutherland et al., 2012], and the unknown, but probably lower geothermal gradient on the southern Alpine Fault, additional experiments are needed to quantify the frictional strength, rate and state friction parameters, and healing behavior of Alpine Fault PSS gouges at conditions experienced by the fault from the surface to the brittle-ductile transition. Observations from this study highlight the importance of considering the cooperative behavior of all parts of an upper crustal fault zone.

2.8 Conclusion

1. Slip on the southern Alpine Fault is localized to a single 1–12 m-thick zone of impermeable, frictionally weak, velocity-strengthening, saponite-lizardite and illite-chlorite-rich foliated gouges. At the surface, coseismic slip appears to propagate along slickensided fault core margins.

2. Australian plate rocks were mylonitized before 70 Ma; thus while the southern Alpine Fault follows a pre-existing locus of deep-seated continental shear, its motion since c. 24 Ma has been insufficient to exhume mylonites derived from the (strike-slip predominant) southern portion of the fault. In contrast, c. 18 Ma Pacific plate mylonites that outcrop here are some of the oldest and southernmost Alpine Fault-related mylonites exhumed by the dextral-reverse central Alpine Fault and subsequently translated to the south by strike-slip motion. All fault-related damage overprinting Pacific plate mylonites here is thus attributed to Neogene slip on the Alpine Fault.

3. The frictionally weakest, velocity strengthening fault core material is associated with a serpentine-bearing mélange that is no older than c. 24 Ma (i.e., is Alpine Fault-related), but possibly younger than c. 13–8 Ma.

4. While the mineralogical and frictional data of Alpine Fault core gouges examined in this study are comparable to those of other major weak-cored faults (e.g., San Andreas Fault, Gokasho-Arashima Tectonic Line), they are incompatible with abundant evidence indicating that the southern Alpine Fault fails in large magnitude quasi-periodic earthquakes. This study highlights the limitations of using surficial outcrop materials and shallow surface, low velocity (below coseismic slip rates of c. 1 m/s) experimental conditions to understand spatio-temporal variations in fault properties.
Chapter 3  Scale Dependence of Oblique Plate Boundary Partitioning: New Insights from LiDAR, Central Alpine Fault, New Zealand

Figure 3.0  Oblique aerial photograph looking northeast along-strike of the Alpine Fault at the western rangefront of the Southern Alps from above the southern extent of the recently-acquired airborne light detection and ranging (LiDAR) survey. Here the active traces of the Alpine Fault zig-zag along the base of the rangefront (i.e., are serially-partitioned into shallowly rooted strike-slip and oblique-thrust segments) at the scale of the major river valleys exiting the Southern Alps. Docherty Creek is in the foreground, Waiho River is at center, and Waitangi-taona and Whataroa rivers are beyond. The LiDAR data successfully reveals surface expressions of the Alpine Fault previously obscured by dense rainforest and indicates parallel-partitioned fault wedges are a widespread feature superimposed on oblique-thrust serial partitions in the region. The town of Franz Josef (at center) straddles a known trace of the Alpine Fault; the width of the Alpine Fault surface rupture zone revealed by the LiDAR data is on the order of 300 m, indicating a significant hazard to the town. Photo taken by L. Homer (CN3760B). Photo courtesy of GNS Science.

This chapter is a version of a peer-reviewed publication: Barth, N.C., V.T. Toy, R.M. Langridge, and R.J. Norris (2012). Scale dependence of oblique plate-boundary partitioning: New insights from LiDAR, central Alpine fault, New Zealand, Lithosphere, 14 p., doi:10.1130/L201.1. Details of the various author contributions are given in Chapter 1. See also Supplement 3 for additional data and discussions.
Abstract

Recently-acquired airborne light detection and ranging (LiDAR) data along a portion of the Alpine Fault are combined with previous work to define how the plate boundary structures partition at three different scales from $< 10^6$–$10^9$ m. At the first-order ($< 10^6$–$10^4$ m), the Alpine Fault is a remarkably straight and unpartitioned structure controlled by inherited and active weakening processes at depth. At the second-order ($10^4$–$10^3$ m), motion is serially-partitioned in the upper ~1–2 km onto oblique-thrust and strike-slip fault segments that arise at the scale of major river valleys due to stress perturbations from hangingwall topographic variations and river incision destabilizing the hangingwall critical wedge, both concepts proposed by previous workers. The resolution of the LiDAR data refines second-order mapping and reveals for the first time that at a third-order ($10^3$–$10^0$ m) the fault is parallel-partitioned into asymmetric positive flower structures, or fault wedges, in the hangingwall. These fault wedges are bounded by dextral-normal and dextral-thrust faults rooted at shallow depths ($< 600$ m) on a planar, moderately southeast-dipping, dextral-reverse fault plane. The fault wedges have widths of ~300 m and are bounded by and contain kinematically stable fault traces that define a surface rupture hazard zone. Newly-discovered anticlinal ridges between fault traces indicate a component of shallow shortening within the fault wedge is accommodated through folding. A fault kinematic analysis predicts the fault trace orientations observed and indicates third-order fault trace locations and kinematics arise independently of topographic controls. A slip stability analysis suggests that new strike-slip faults will easily accommodate displacement within the hangingwall wedge, and that thrust motion is most easily accommodated on faults oblique to the overall strike of the Alpine Fault. This study suggests the thickness of footwall sediments and width of the fault damage zone (i.e., presence of weaker, more isotropic materials) are major factors in defining the width, extent and geometry of third-order near-surface fault wedges.

3.1 Introduction

In the last 10 years, airborne light detection and ranging (LiDAR) technology has been demonstrated to be an effective tool to identify active faults and assess seismic hazards in a wide range of environments including urban areas (northern Taiwan: Chan et al., 2007; Houston, Texas: Enhelkemeir and Khan, 2008), grassland and scrub (California: Arrowsmith and Zielke, 2009; Zielke et al., 2010; Hunter et al., 2011) and densely-vegetated forests (Washington: Harding and Berghoff, 2000; Hauegerud et al., 2003; California: Zachariasen and Prentice, 2008; Slovenia: Cunningham et al., 2006). In particular, the ability of LiDAR to image the land surface
beneath thick vegetative covers at a resolution of centimeters to meters has allowed the
topographic features of these landscapes to be examined in revolutionary new ways.

Airborne LiDAR data were collected in July 2010 in a 1.5 km-wide by 34 km-long swath
encompassing a portion of the heavily-vegetated Alpine Fault (the major plate-boundary
structure in the South Island of New Zealand, Figure 3.1) to better understand seismic
hazards in the region. The LiDAR survey extends roughly between the townships of Franz
Josef (which straddles a known surface trace of the Alpine Fault) and Whataroa. Within the
study area approximately 75% of the terrain is covered by dense temperate rainforest, with
the remaining 25% dominated by lightly-populated active river floodplains. Within the latter,
any evidence for a surface rupture of the most recent Alpine Fault earthquake in AD 1717
[Yetton, 1998; Yetton et al., 1998] is already erased. The prospect of a 30% chance of a surface
rupturing ~M8 Alpine Fault earthquake in the next 50 years [Sutherland et al., 2007a; Berryman
et al., 2012a] poses a significant national hazard to the country of New Zealand, and better
understanding of surface rupturing along the fault is sorely needed. By incorporating
interpretations of the new LiDAR data with aerial photo interpretation and decades of
previous geologic mapping, I explore the significance of several orders of magnitude of
shallow transpressional partitioning observed on the oblique-slip central Alpine Fault and
propose a new model for the geometry of structures observed at 10^3–10^0 m scales.

3.2 Geologic and Tectonic Setting

The Alpine Fault is the ~880 km-long (~600 km onshore) major active transpressive plate
boundary structure in the South Island of New Zealand (Figure 3.1). It currently
accommodates ~75% of total motion between the Pacific (PAC) and Australian (AUS) plates
[Berryman et al., 1992; Norris and Cooper, 2001; Sutherland et al., 2006; Barnes, 2009; Langridge et al.,
2010]. Along with the North Anatolian Fault in Turkey and the San Andreas Fault in
California, the Alpine Fault is commonly cited as one of the major continental strike-slip
faults in the world [e.g., Sylvester, 1988; Molnar and Dayem, 2010; Frankel and Owen, 2013].

In the study area the Alpine Fault is the geologic boundary between granitic and gneissic
Western Province (AUS plate) and gneissic and schistose rocks (Alpine Schist and Alpine
Schist-derived mylonites) of the PAC plate [Reed, 1964; Cox and Barrell, 2007] across which
480 km of cumulative strike-slip motion has taken place [Wellman, 1949 in Benson, 1952]. Dip-
slip rates vary along-strike of the Alpine Fault. Only its central part experiences very high
Figure 3.1  Tectonic setting of New Zealand showing major plate boundary structures and LiDAR study area overlain on bathymetric and topographic data compiled by the National Institute of Water and Atmospheric Research of New Zealand (NIWA). Coordinates in decimal latitude and longitude.

hangingwall uplift rates, giving rise to the Southern Alps. In the vicinity of the LiDAR study area, the fault is a relatively planar, moderately southeast-dipping dextral-reverse fault accommodating ~27 mm/yr of strike-slip motion and up to ~10 mm/yr of dip-slip motion [Norris and Cooper, 2001; Little et al., 2005; Sutherland et al., 2006]. The fault is thought to rupture in large to great (~M 8) earthquakes with a mean long-term recurrence interval of c. 300 years and last ruptured in AD 1717 [Yetton, 1998; Yetton et al., 1998; Wells et al., 1999; Sutherland et al., 2007a; Berryman et al., 2012a]. In terms of single-event displacement, Alpine Fault earthquake events have generated 6–9 m of dextral displacement and a throw of up to 2 m [Cooper and Norris, 1995; Yetton and Nobes, 1998; Berryman et al., 2012b]. Evidence from fault trenches and tree-ring chronologies suggest that at least 375 km strike-length of the fault ruptured in the AD 1717 earthquake [Wells et al., 1999].

The Alpine Fault forms a particularly striking lineament visible from space, extending for hundreds of kilometres (Figure 3.2A). While much of the fault has a relatively straight and uncomplicated surface expression at this scale, the central section exhibits sequences of transpressional oblique-thrust and strike-slip faults (Figure 3.2B) dubbed “serial partitioning”
Figure 3.2  (A) Extent of the onshore trace of the Alpine Fault (arrows) visible from topography. Notice the straightness of the Alpine Fault at this scale (< $10^6$–$10^4$ m). All images in this figure utilize a hillshade derived from the Land Information New Zealand 100 m DEM. Illumination is from the northwest. All coordinates in decimal latitude and longitude. (B) Major serially-partitioned (oblique-) thrust and strike-slip faults mapped from field, aerial photographic and LiDAR data [refined after Norris and Cooper, 1995]. Serially-partitioned segments are of the order of $10^4$–$10^3$ m in length. The strike-slip faults occur at the scale of the major river valleys and can be thought of as passive linkage structures necessitated by tears in the basal fault plane in the near-surface. (C) All fault traces identified in this study ($n = 268$) mapped at $10^3$–$10^0$ m scale from LiDAR data, field checking and previous work. Local AUS-PAC plate motion vector from NUVEL-1A [DeMets et al., 1994]. Towns of Franz Josef and Whataroa are denoted in capital letters. Location names are those referred to in Figure 3.4. 

by Norris and Cooper [1997; 1995]. These zig-zag sequences of north-northeast-striking thrust/dextral-thrust faults and east-northeast-striking dextral faults repeated over lengths of 1–10 km have been interpreted to form as a result of an angle of obliquity of 16° between the 071°-trending relative PAC-AUS plate motion vector [DeMets et al., 1994] and the 055°-striking central Alpine Fault, as well as local stress field perturbations due to topographic effects of a steep range-front with deeply-incised river valleys [Norris and Cooper, 1995, 1997].
Norris and Cooper [1995] constructed transpressional sandbox experiments simulating the Alpine Fault with and without valleys to support the topographic influence on serially-partitioned fault segments. Serial partitioning here is confined to the upper ~1–2 km of the crust along approximately 100 km of the central Alpine Fault.

An alternate geometry, parallel partitioning, in which oblique motion is accommodated on parallel-striking thrust and strike-slip faults, is more common globally [e.g., Wentworth and Zoback, 1989; McCaffrey, 1992, 1996; Molnar and Dayem, 2010]. Parallel partitioning has been documented in detail on the Alpine Fault offshore Fiordland by Barnes et al. [2005] where multiple fault splays ~1–2 km apart are interpreted to link into a single throughgoing fault at depth. At Darnley and Gaunt Creeks in the LiDAR study area, structural mapping by Easterbrook [2010] revealed two subvertical hangingwall dextral faults thought to root into a southeast-dipping dextral-thrust fault at shallow depths to form a ~200 m-wide zone of parallel partitioning. These two examples in Easterbrook’s study are the first suggestion of parallel partitioning on the central Alpine Fault in the same area where serial partitioning was previously identified. A requirement of serial partitioning is that equal amounts of slip are accommodated on sequenced thrust and strike-slip fault segments, whereas more variable slip partitioning is possible between parallel-partitioned thrust and strike-slip faults.

Imbricate out-of-sequence thrusts form as rapid river incision reduces the wedge taper below its critical value so that they internally imbricate [Norris and Cooper, 1997]. The best known example of this is the Waikukupa River-Hare Mare Creek thrust system where the wedges can be > 1 km-wide. A 30° dipping basal plane requires a negative surface slope to render the wedge subcritical and therefore likely to imbricate in this way [Norris and Cooper, 1997]. Negative surface slopes (i.e., towards the range) are only likely to occur where deep incision by rivers has taken place, hence the close relationship to valleys. Other prominent examples of abandoned thrusts of the Alpine Fault are at Ward Hill near Paringa 60 km, southwest of the study area [Simpson et al., 1994], and near Simon Slew and Robinson Creek, 85 km southwest of the study area [Department of Geology, University of Otago, 2012, Alpine Fault map Web site].

Despite the complex surface rupture patterns, all evidence suggests the fault is a single planar structure at depths greater than 500–1000 m and that shallow partitioning does not appear to hinder the propagation of earthquakes through this region [Norris and Cooper, 1995]. Along its entire onshore length, the fault exhibits no step-overs or discontinuities in the fault plane greater than 1 km, at depth [Sutherland et al., 2007a]. Active faults within the Western Province
are rare in the central Alpine Fault zone [Cox and Barrell, 2007]. As the Alpine Fault dips shallowly to moderately southeast with depth, fault traces south and east of the Alpine Fault are rooted within the base of the PAC plate (the main Alpine Fault plane). Thus the main trace of the Alpine Fault tends to correspond to the most basin-ward (basal) fault trace, which still accommodates the most slip at the surface, with partitioning occurring only in the hangingwall.

Typical exposures along the central Alpine Fault (e.g., Hare Mare Creek- 4 km southwest of study area, Gaunt Creek) consist of PAC plate fault rocks thrust obliquely over fluvio-glacial sediments on the AUS plate. The PAC plate section typically comprises a cm-thin ultracataclastic clay gouge, a 20–30 m-thick hydrothermally-altered highly-fractured fault rock (commonly referred to as “cataclasite”) and a sequence of progressively less-fractured ultramylonite, mylonite, protomylonite and paragneissic Alpine Schist [Wellman, 1955a; Sibson et al., 1981; Norris and Cooper, 2007]. The zone of intense fracturing (i.e., damage zone) associated with the Alpine Fault is typically on the order of 100 m in width on the PAC plate [e.g., Cooper and Norris, 1994; Norris and Cooper, 1997; Wright, 1998a]. These rocks are especially prone to near-surface faulting and slope failure [Korb, 2004].

The Alpine Fault outcrops at the foot of the steep western range-front of the Southern Alps. Adjacent to the high uplift region around Franz Josef are some of the highest peaks (~3000 m) along the main divide of the Southern Alps; these are found at a distance of about 15 km southeast of the surface trace of the Alpine Fault. Fast-flowing mountain rivers in steep-walled glaciated river valleys exit the Southern Alps and cross the Alpine Fault at elevations of ~100 m onto wide and active river floodplains lined with widespread glacial moraine deposits attributed to as many as five major Middle to Late Pleistocene glacial advances [Barrell, 2011]. The most recent widespread glacial advance dates to the early or mid-Last Glacial Maximum (LGM), c. 27,000–20,000 years BP, which covered most of the range-front of the Southern Alps in ice [Suggate, 1990; Cox and Barrell, 2007; Barrell, 2011]. A high annual precipitation of 5000 mm allows a dense temperate rainforest to flourish west of the main divide.

3.3 Methods

LiDAR data used in this study were collected by NZ Aerial Mapping Limited using an Optech ALTM3100EA instrument flown from ~1200 m altitude. Data were collected in six overlapping swaths ~780 m wide (for a total width of ~1.5 km) to increase the point counts
through the dense podocarp rainforest. Positional accuracy of data points is ± 0.30 m horizontally and ± 0.15 m vertically. A 2 m resolution bare-earth digital elevation model (DEM), 2 m hillshade, and 0.5 m contour lines generated from the last returns (i.e., ground returns) of the LiDAR data were used in this study. These data were incorporated into a geographic information system (GIS) geodatabase along with Land Information New Zealand (LINZ) Topo 50 topographic map sheets, and georeferenced geologic maps and structural data collected previously by researchers from the University of Otago and GNS Science [Cox and Barrell, 2007]. Orthophotos taken at the same time as the LiDAR run were also incorporated into the geodatabase and were most helpful for identifying linear cultural features, such as roads and trails, which could be mistaken for natural features. ArcMap computer software was used to create polyline and polygon shapefiles of features of interest including fault traces, anticlinal ridges, lineaments, fault offset features, landslides, river terraces, and glacial features. Interpretation was initially performed without reference to previous geologic information, using only the LiDAR data and orthophotos, to eliminate potential bias.

Special care was taken in the interpretation of fault traces, especially since regional fracture patterns and Alpine Schist/mylonite foliations are known to have similar strikes to expected fault traces [Hanson et al., 1990]. Low-angle range-front dextral-thrusts, which commonly form the most basin-ward Alpine Fault trace (e.g., Waikukupa Thrust, Hare Mare Creek, Gaunt Creek), consist of highly-fractured fault rocks and incoherent fluvio-glacial deposits thrust over unconsolidated fluvio-glacial gravel deposits. These fault traces can be subtle features due to the low angle at which the fault plane intersects topography, and are commonly obscured by deposits from slope failures prompted by tectonic over-steepening, fluvial undercutting and earthquake ground motion in highly-fractured rock. Dextral-thrust fault traces are characterized by 20–40 m-high over-steepened range-front topographic steps onto river flood plains or alluvial fans. In contrast, high-angle dextral and dextral-normal faults have well-defined traces due to the high angle at which the fault plane intersects topography and relative protection from fluvial erosion in the interfluves. These fault traces are characterized by sharp linear or curvilinear scarps 1–10 m-high, or symmetrical troughs. While the strike-slip-dominant fault scarps are the most striking, it is emphasized that they have not necessarily accommodated the most movement or are a product of the youngest surface ruptures.
The features were sorted into several shapefiles utilizing a hierarchy of confidence in interpretation of fault traces. Factors such as definitive offsets, adjacency to anticlinal ridges, scarps on millennially-inactive glacial or fluvial surfaces, particularly well-defined range-front lineaments, signs of disequilibrium erosion, and otherwise atypical geomorphology best explained by the presence of a fault were used to qualitatively rank individual features as “fault trace,” “probable fault trace” or “lineament” (in order of decreasing confidence). After several iterations, comparisons with previous work and some field checking, this classification was simplified into fault trace and lineament shapefiles. An effort was also made to define the primary slip zone; the fault trace (or traces), which has accommodated the majority of surface rupture slip at any point along-strike.

3.4 Results

3.4.1 Fault traces

In total, 268 fault traces were identified within the LiDAR coverage area in a 200–800 m-wide zone along the range-front (Figure 3.2C) for a total fault trace length of 68.9 km. I have compared my results to a compilation of Alpine Fault fault traces mapped by researchers at the University of Otago’s Department of Geology (publicly available through their website). I found 12.4 km of the total fault length in this study had been previously identified as “accurate” (±50 m) and that by length 82% of the LiDAR-mapped traces are new. Only 4.2 km (6%) of the fault traces southeast of the range-front were previously identified.

Mean fault trace length is 260 m with a few traces longer than 1 km. Many ends of fault traces are covered or have been removed by slope failure or creek erosion and aggradation, suggesting the fault traces would have been longer and more continuous at the time they ruptured. Some fault traces extend off the LiDAR-mapped area, suggesting there are places where extending the coverage further southeast may reveal more fault traces. However, in most places the LiDAR imagery covers the full width of surficial structures I infer are part of the Alpine Fault surface damage zone. I observe no fault traces within the interfluvues northwest of the range-front and that fault traces tend to occur at low elevations north or west of the steepest portion of the range-front step. Fault mapping across the whole width of the central Southern Alps by Cox et al. [2012a] also shows a relative absence of faults 5–10 km southeast of the study area. Field investigations west of the Whataroa River have allowed me to confirm that LiDAR-mapped fault traces are indeed faults, and have also allowed me to identify 1–2 m-wide linear troughs in Riedel shear orientations below the resolution of the
LiDAR data. In addition, trenching of a north-striking fault trace at Gaunt Creek confirmed the existence of a multi-event thrust fault [De Pascale and Langridge, 2012].

I recognize eight $10^4$–$10^3$ m-scale serially-partitioned segments in the LiDAR coverage area. The mapped segments are largely concordant with the regional-scale serial-partitioned segments mapped by Norris and Cooper [1995] with the exception of a new well-defined 3.5 km-long strike-slip segment between Docherty Creek and Franz Josef Township (Figure 3.2B). My mapping also shows a clearer link between serial-partitioned segments and major river valleys than these authors were able to make previously; the south-western ends of oblique-thrust segments correlate to the south-western side of major river valleys (e.g., Docherty Creek, Waiho River, Waitangi-taona River, Whataroa River). The overall strikes of oblique-thrust segments are 030–053° (048° average) and strike-slip segments are 068–088° (082° average). Thrust segments tend to be longer, although the longest strike-slip segment south of Whataroa Township is ~5 km-long.

A rose diagram of $10^3$–$10^0$ m-scale fault traces weighted by fault length gives thrust and dextral-thrust maxima at 040–045°, oblique maxima at 060–065°, and dextral maxima at 070–075° (Figure 3.3); when combined with structural measurements these maxima are interpreted as thrust/dextral-thrust, oblique, and dextral fault sets respectively. Range-front dextral-thrust fault traces are typically curvilinear and are expressed as poorly-preserved west-to-northwest-facing 20 m high scarps. Dextral-normal fault traces within 100–1000 m of the range-front are expressed as linear or curvilinear 1–10 m-high south-to-southeast-facing scarps or 10 m-wide symmetrical troughs. These dextral-normal traces have not been identified prior to this study, but are thought to be coincident with east-west striking clay gouge-cored faults observed in Docherty Creek [e.g., Norris and Cooper, 1995], which dip 70–90°S with groove lineations raking 20–60° from the west indicating dextral-normal slip.

Dextral stream channel offsets are typically ambiguous along the range-front dextral-thrust traces. Alluvial fans have a tendency to deflect streams to the southwest (apparently sinistrally) as the highest part of the fan is shifted to the northeast relative to the mouth of the valley. The more range-ward faults have identifiable dextral offsets of creeks, particularly in the area west of the Whataroa River and near Darnley Creek, in part due to the creeks having deeply-incised channels on both sides of fault traces. I find the LiDAR study area illustrates the complications in determining offset rates on faults with multiple rupture planes in the near-surface. It is common to observe several parallel major traces; these may have been active at different times and may have accommodated different components of strike-
slip and dip-slip motion. Thus a slip rate determined from one of these traces would not be representative of the entire rupture history.

![Rose diagram of 268 fault trace azimuths weighted by length with 5° binning. Total fault trace length is 68.9 km. Maxima are interpreted from structural measurements. Arrow denotes average strike of the Alpine Fault (055°).](image)

**Figure 3.3**

3.4.2 Characteristic fault geometries

Characteristic transpressional fault geometries observed in this study are highlighted in Figure 3.4. McCulloughs Creek and Darnley Creek (Figure 3.4A–B) are both examples of parallel partitioning in which a 100–300 m-wide fault wedge is bounded by a basin-ward oblique-thrust and a range-ward dextral-normal fault rooted on a southeast-dipping Alpine Fault plane. At McCulloughs Creek the terrace riser north of the creek is offset about 100 m dextrally and 10 m down-to-the-southeast along the dextral-normal fault. A complex array of normal faults lie in the southeastern half of the fault wedge while the northwestern half is dominated by three en echelon anticlinal ridges clearly defined from LiDAR-derived contour data that comprise the axis of the anticlinal fault wedge. At Darnley Creek in the Waitangi-Taona Valley, a curvilinear thrust trace and a linear 10–25 m down-to-the-southeast dextral-normal trace bound the fault wedge. To the north the dextral-normal fault offsets two stream
channels whereas to the south it devolves into 75 m-long en echelon dextral-normal traces. Anticlinal ridges are parallel to the trace of the main dextral-normal fault.

Figure 3.4 Key examples of characteristic geometries revealed by LiDAR data. Fault motion is inferred based on LiDAR data and limited field data. Fault and fold interpretations overlain on 2 m LiDAR hillshade. Grid North is towards the top of the figure in all views. Refer to Figure 3.2C for locations. (A) McCulloughs Creek and (B) Darnley Creek sites illustrate third-order parallel partitioning on a second-order thrust segment. (C) The Franz Josef area is characterized by en echelon fault partitioning. (D) Arthur Creek illustrates fault patterns on a second-order strike-slip segment. (E) Docherty Creek –Waiho River and (F) Gaunt Creek–Matainui Creek are on transitions from thrust and strike-slip segments.
The region between Franz Josef Township and Tatare Stream (Figure 3.4C) is characterized by en echelon strike-slip faults that curve into parallelism with the rear of the fault wedge such that the overall character is similar to parallel partitioning. The basin-ward side of the fault wedge is dominated by oblique-thrust faults and the range-ward side by dextral-normal faults. The range-front thrust faults have small apparent dextral offsets where they cross two small drainages, which correspond to dextral faults. This behavior is comparable to that generated in sandbox experiments conducted by Norris and Cooper [1995] using a rigid indenter and ignoring the effects of topography. The dextral faults have increasing down-to-the-southeast displacement as the faults approach parallelism with the rear of the fault wedge.

At Arthur Creek two major fault traces are observed dextrally offsetting stream channels on the longest strike-slip serially-partitioned segment in the study area. Well-defined east-northeast-striking anticlinal ridges (compared to more northeasterly-striking anticlinal ridges on thrust segments) accommodated limited shortening on the basin-ward side of fault traces. No thrust traces were identified here. Fault traces here record the most cumulative dextral displacement preserved from offset features anywhere in the LiDAR study area.

The Docherty Creek–Waiho River and Gaunt Creek–Matainui Creek regions (Figure 3.4E–F) are on transitions between near-orthogonally striking thrust and strike-slip serially-partitioned segments. From Docherty towards the Waiho, northeast-striking thrust faults transition to east-northeast-striking dextral-thrust faults to east-striking dextral-normal faults. Thrust fault traces still bound the basin-ward side of the fault wedge on this strike-slip segment. From Gaunt Creek towards Matainui Creek, north-northeast-striking thrust faults transition to east-striking dextral faults along the range-front. More range-ward dextral faults strike northeast, effectively “cutting the corner.” Note that the width of the fault wedge (and density of faults) is greatest at the serially-partitioned transition.

### 3.4.3 Lineaments

In addition to fault traces, 175 lineaments were identified (Figure 3.5). Most of these are more range-ward than the fault traces and cut topography steeply, frequently as strikingly linear drainages. I found lineaments (100–500 m-long) revealed in the LiDAR imagery within 1 km of the Alpine Fault have similar orientations and abundances to large-scale regional lineaments within 7.5 km east of the Alpine Fault near Franz Josef and Fox Glacier compiled from aerial photographs by Hanson et al. [1990]. Hanson et al. [1990] identified 3 orientation
sets of lineaments: a 025°-trending set parallel to the strike of the foliation in the Alpine Schist, a minor set trending 130° possibly related to extension fractures and an east-west trending set of dextral faults with late normal displacement. The structures in both studies are similar in orientation despite the fact that the bulk of the area examined by Hanson et al. [1990] was in the Alpine Schist, which is the protolith of the Alpine Fault mylonites that underlie most of the study area described herein. I find structures parallel to the Alpine Fault and mylonitic foliation are more dominant in my dataset than in Hanson et al. [1990] (Figure 3.4). The dominant strike (035°) is slightly more parallel to the Alpine Fault than was observed by Hanson et al. [1990], likely indicating a clockwise rotation towards parallelism with the Alpine Fault with decreasing distance from the fault. This is consistent with dextral shear having been accommodated in the ductile shear zone at depth with concurrent foliation development. For shear strains > 5 as measured by Norris and Cooper [2003], the angle between foliation and the shear zone should be < 10°.

Figure 3.5 Lineaments (excludes fault traces). (A) Rose diagram (this study) showing azimuths of 175 lineaments with 5° binning. (B) Rose diagram redrawn from Hanson et al. [1990] showing the azimuths of 788 lineaments with 10° binning. Lineaments from Hanson et al. were traced from aerial photography of an area bounded by the Alpine Fault, the Fox and Waiho Rivers, and ~7.5 km east of the Alpine Fault.
3.4.4 Anticlinal ridges

Anticlinal ridges associated with near-surface Alpine Fault deformation have not been previously identified within the study area. I map 100 anticlinal ridges with characteristic strikes 0–15° counter-clockwise to the nearest fault trace strike (Figure 3.6). The resolution of the LiDAR reveals that anticlinal ridges are composed of smaller scale en echelon anticlines, which could indicate development in a stepwise manner (e.g., Figure 3.4A). The majority of anticlinal ridges cluster around a strike of 055° (the average strike of the Alpine Fault), with a minority approaching a strike of 085°, close to the average strike of strike-slip fault traces in this study. In the field I observed one of these anticlinal ridges near Arthur Creek, west of the Whataroa River (Figure 3.4D). It comprises Whataroa-derived bedded river terrace deposits folded into an asymmetric anticline with a moderately-dipping northwest limb and shallowly-dipping southeast limb. A comparable recently-uplifted asymmetric anticlinal structure associated with hangingwall near-surface Alpine Fault deformation has been identified near Paringa about 60 km south of the study area [Suggate, 1968; Adams, 1979; Simpson et al., 1994].

Figure 3.6    Rose diagram of 100 anticlinal ridge axes with 5° binning.
3.4.5 Fault kinematic analysis

I use field data to predict dips for the lineaments identified as fault traces. The hangingwall sequence of fault rock, comprising mylonites and cataclasites developed during ductile then brittle shear at depth on the Alpine Fault, is exposed in stream sections to the southeast of the most recent trace [e.g., Toy et al., 2008; Norris and Cooper, 2007]. Uncemented clay gouge zones outcrop throughout this exhumed sequence [Sibson et al., 1979; Reed, 1964], including at the contact of hangingwall cataclasite on footwall gravel. I assume the gouge zones developed during near-surface fault slip since they have not been affected by hydrothermal cementation. Their mean orientation (Figure 3.7) is 030°/40°SE, but dips ranging from 30–50° are common. Additionally, the mylonitic foliation in the hangingwall, which has an average orientation of 055°/50°SE [Norris and Cooper, 2007] presents a strong mechanical anisotropy and brittle faults should therefore parallel it. The dominant orientation of the brittle part of the fault at depth (below near-surface complexities) can be explained in this way.

![Figure 3.7](image)

**Figure 3.7** Equal area lower hemisphere projections illustrating (A) poles to gouge zones measured within the central Alpine Fault zone (black dots); Kamb contour interval = 2.0, n = 224; (B) orientations of faults interpreted from combined LiDAR and field data (black great circles); slip vector for three of the fault sets indicated by grey dot; (C) Expected orientations of faults in a Riedel shear system around a mean Alpine Fault plane (Y-shear) orientation of 055°/45°SE. Dashed great circle indicates hinge plane of associated folds. Arrows and D/U show shear sense. Arrow outside plot shows trend of plate motion vector. In (B) and (C), grey bars outside circle illustrate strikes of main groups of faults interpreted from the LiDAR data.

However, in the ~10 well-exposed thrust segments of the fault, the boundary between hangingwall cataclasite and footwall gravel commonly dips more shallowly (~30°) than the mylonites that overlie it. This may be due in part to near-surface stress control as this shallower orientation has a dip similar to that predicted for thrust faults in the near surface, where a principal stress is likely to be vertical [Anderson, 1951]. It may also be related to
collapse of the overthrust material toward the free surface. By the same principle, I expect strike-slip faults to be near vertical. Reassuringly, steeply S-dipping, E-W striking clay gouge zones with (sub)-horizontal striae have been observed in a number of places in the hangingwall sequence (Figure 3.6A) [Darnley Creek: *Easterbrook*, 2010; Gaunt Creek: *Toy*, 2007; Docherty Creek and Waikukupa River: *Norris and Cooper*, 1995; 1997]. These vertical features are easily eroded in creek sections and do not outcrop as commonly as thrusts, so are under-represented in the compilation of gouge zone orientations (Figure 3.7A).

By combining fault strikes from LiDAR with corresponding field observations of characteristic fault dips and striae on faults, I hypothesize a set of likely orientations and slip vectors for the major groups of faults identified in the LiDAR data (Table 3.1). If faults of these orientations bound semi-rigid crustal blocks, their intersections must approximately parallel the fault slip vectors in order to maintain geometrical continuity. For a slip vector parallel to the NUVEL-1A plate motion vector trending 071° [*DeMets et al.*, 1994], some fault orientations are thus further constrained so a single slip vector orientation (15° → 071°) can lie within them, as indicated in Table 3.1 and Figure 3.7B. From Figure 3.7B it is also apparent that it is difficult for such geometrical continuity to be maintained during coeval slip on reverse faults and faults of any other type. This may induce the fault to partition into the other orientations as it approaches the free surface. I present a three-dimensional model for the geometry of interlinked fault segments with these orientations in the Discussion.

**Table 3.1** Orientations of fault groups evident in LiDAR data, with dips constrained by field observations as stated in the text.

<table>
<thead>
<tr>
<th>Orientation of fault plane (strike/dip)</th>
<th>Orientation of slip vector (plunge/trend)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Major thrust segments; near surface</td>
<td>040–045°/30°SE</td>
</tr>
<tr>
<td>Major reverse segments at greater depths</td>
<td>040–045°/50°SE</td>
</tr>
<tr>
<td>Major strike slip segments</td>
<td>070–075°/90°</td>
</tr>
<tr>
<td>Oblique segments</td>
<td>060–065°/60°SE</td>
</tr>
</tbody>
</table>
An alternate model for likely fault orientations can be constructed based on the idea of a characteristic Riedel shear geometry. In natural situations, the elastic stress field generated in the material surrounding a slipping fault, particularly within a relatively unconsolidated (structurally weak) sequence above a basement fault, results in formation of a system of linked faults with characteristic geometry \([\text{Cloos}, 1928; \text{Riedel}, 1929]\). Using this characteristic fault geometry, for a main fault orientation of 055°/45°SE \([\text{Norris and Cooper}, 2007]\) and a slip vector parallel to the \(\text{DeMets et al.}[1994]\) PAC-AUS plate motion vector, Figure 3.7C shows a predicted fault system geometry. Most of the subsidiary fault orientations in this case strike parallel to fault traces identified in the LiDAR (e.g., P-shear at 038°/52°SE is similar to the group of faults with strikes 040–045°; R-shear at 075°/42°SE is similar to the group of faults with strikes 070–075°). Therefore, many of the faults identified in the LiDAR could have initiated as part of a Riedel shear system arising independently of topographic and river incision effects. However, this system of faults cannot intersect parallel to the slip vector, so is geometrically destabilized by ongoing slip. Internal deformation of Riedel shear-bounded blocks by folding could allow shear to continue on these structures. In either case, it is important to realize that displacement within the near-surface fault zone is controlled by slip/creep on the deep fault zone and so the surface structures are constrained to an average strike similar to the deep regional structure and are required to accommodate oblique motion.

3.4.6 Fault stability and slip tendency analysis

The tendency of any particular fault orientation to slip was evaluated by calculating the relationship between the static coefficient of friction \((\mu_s)\) and the ratio of maximum and
minimum principal stresses ($R = \sigma_2 / \sigma_3$) at which slip will occur. This is accomplished by resolving the components of shear stress ($\tau$) and normal stress ($\sigma_n$) on the fault plane for various $R$, then applying the limiting condition for failure $\tau/\sigma_n > \mu_s$ [Collettini and Trippetta, 2007] for a range of $\mu_s$. This calculation was performed iteratively for increasing values of $R$ until this limiting condition was met, using a simple MATLAB script. Note that it is assumed that $\sigma_2 = \frac{1}{2} (\sigma_1 + \sigma_3)$ but similar results are obtained if $\sigma_2 \approx \sigma_3$.

The results (Figure 3.8) indicate that steeply-dipping (i.e., strike-slip) faults have a greater tendency to slip than other orientations indicated in Table 3.1. This is apparent from the fact that, for a comparable ratio of principal stresses, a strike-slip fault will slip even if its frictional strength ($\mu_s$) is over twice that of any of the predicted thrust, or oblique-thrust fault orientations. On the other hand, there is little difference in the strength of the various thrust and oblique-thrust fault orientations indicated by the LiDAR analysis when expressed in this way. For comparison purposes only, a fault with the average strike of the Alpine Fault overall, and dip parallel to the exhumed mylonitic foliation (055°/45°SE) was also evaluated. This structure has even lower tendency to slip than any of the structures derived from the LiDAR analysis. This result suggests that new strike-slip faults will easily accommodate displacement within the hangingwall wedge, and that thrust motion is most easily accommodated on faults oblique to the overall strike of the Alpine Fault. The relative tendency of these fault orientations to slip is consistent with the formation of a partitioned fault system in the near-surface to accommodate oblique motion away from a non-ideally oriented fault plane, and with observations that fault traces in this study are reactivated during multiple earthquake rupture events.

It is unlikely that the local stress tensor around the faults is the same as that derived at depth by Leitner et al. [2001] and Boese et al. [2012] due to variations in topography and therefore in the orientation of the free surface. The anisotropic strength of the strongly-foliated mylonites will also influence fault orientations and slip tendency [e.g., Donath, 1964]. Furthermore, it is possible that the stress tensor is perturbed by the existing faults, so that a heterogeneous stress distribution allows slip on all fault orientations, despite their varying strengths. Essentially kinematic constraints mean that slip has to occur on these faults and stresses will locally rotate to allow this. Small-scale stress variations should be more thoroughly evaluated in the future by combining numerical models with in situ measurements derived from future borehole measurements.
3.5 Discussion

3.5.1 First-order controls on the Alpine Fault

The Alpine Fault is thought to have localized on an inherited Eocene (c. 45 Ma) passive
margin that separated continental and oceanic lithosphere [Sutherland et al., 2000]. The Eocene
rift boundary exploited Cretaceous oceanic transform faults and a major crustal discontinuity
within the Zealandia continent that could be as old as Paleozoic [Sutherland et al., 2000].

The first-order structure (< 10⁶–10⁴ m) of the central Alpine Fault is thus simply imposed by
the fact it experiences a component of rapid dip-slip motion throughout the continental
upper crust of the South Island along a progressively weakening discontinuity. The fault zone
only displays time-averaged dominantly brittle behaviour in the upper ~8 km [Toy et al., 2010].
Beneath this, viscous creep predominates within a ductile shear zone that continues to the
base of the quartzofeldspathic crust at ~35 km [Little et al., 2005]. Ductile shear results in
development of a > 1 km thick zone of metamorphic tectonites with finer grain size than the protolith [i.e., mylonites; Sibson, 1977]. Total simple shear strains range up to $\gamma = 180–220$, so a planar fabric is well-developed parallel to the shear zone boundaries [Ramsay, 1980; Norris and Cooper, 2003]. Hot hangingwall rock is uplifted relative to the footwall by shear displacement on the fault faster than diffusive cooling can occur, resulting in advection of hangingwall isotherms across the ductile shear zone [Koons, 1987]. This results in thermal weakening and localization of ductile shear, so that both dip-slip and strike-slip components of relative plate motions are focused on a single structure dipping 40–60°SE [Koons et al., 2003; Norris and Cooper, 2007], a reasonably unfavorable orientation for slip according to an Andersonian view of fault mechanics [Anderson, 1951]. Transient high stresses and high strain rates due to loading around a very narrow fault tip at the base of the seismogenic zone contribute to keeping the shear zone localized within the deforming lower crust [Ellis et al., 2006].

The ductilely-sheared rocks are progressively exhumed up dip of the fault zone during successive earthquakes. At depths where viscous creep no longer predominates (< ~8 km), brittle shear remains localized within the exhumed shear zone rocks, parallel to the boundaries of this zone at depth. This occurs because cohesion parallel to the well-developed shear zone boundary-parallel foliation in the micaceous mylonites is lower than in any surrounding rocks. Consequently, brittle Mohr-Coulomb failure is most likely to occur parallel to the existing fabric [cf. Allen and Shaw, 2011]. The end result is a fault zone that fails parallel to the inherited mylonitic foliation, which dips 50–60°SE.

3.5.2 Second-order controls on the central Alpine Fault

At a second-order ($10^4$–$10^3$ m) the central section of the Alpine Fault (where uplift rates are high) is partitioned along-strike into sequenced (i.e., serial) oblique-thrust and strike-slip segments. These arise in the upper < 1–2 km of the crust when stress perturbations from hangingwall topographic variations (a steep range-front with deeply glaciated valleys and high peaks) on a dipping oblique-slip structure exceed the strength contrast imposed by the anisotropic tectonite fabric [Norris and Cooper, 1995, 1997]. My mapping illustrates a clear link between serial-partitioned segments and major river valleys with the south-western ends of oblique-thrust segments correlating to the south-western side of major river valleys, such as at Docherty Creek, Waiho River, Waitangi-taona River, and Whataroa River. At these same transitions active strike-slip traces do not extend past the oblique-thrust segments into the
hangingwall, indicating that strike-slip movement is limited to the section between the offset oblique-thrusts. This clearly supports the concept of strike-slip segments as linkage structures between (and only between) tears in the basal fault plane. The LiDAR data therefore provide strong support for the basic idea of serial partitioning proposed by Norris and Cooper [1995, 1997]. Slip on the strike-slip segments should thus depend on (and be equal to) slip on the oblique-thrust segments in the manner oceanic transform faults are dependent on oceanic ridge spreading. As such, the strikes of the strike-slip segments would be expected to be close to the azimuth of the plate motion vector accommodated on the Alpine Fault. The average strike-slip segment strike of 082° might locally indicate an 11° clockwise rotation of the slip vector from the 071° plate motion vector, which would be consistent with observations of greater obliquity and higher uplift rates along the central Alpine Fault. Oblique-thrust segments tend to be longer because they are continuous at depth with a 055°-striking dextral-reverse fault plane. The strike-slip segments are tears within this surface geometrically constrained to an average 082° strike. The amount of rotation of the oblique-thrust adjacent to these tears, and hence the length of the strike-slip segment, is related to the amount of westward thrusting, which in turn is constrained by the spacing of major river valleys (~10–15 km) and by the 082° average strike of the tears.

Recent drill core recovered from the Deep Fault Drilling Project (DFDP-1) at Gaunt Creek has revealed that where the basal fault is thrust over gravels at 90 m depth it dips 30° (DFDP-1A). Dip increases to >30° in another hole (DFDP-1B) 100 m away where the hangingwall PAC Plate fault rocks are juxtaposed against the footwall AUS Plate fault rocks at a depth of ≤130 m [Sutherland et al., 2011]. There may be similar dramatic changes in the dip of the fault plane over a short distance elsewhere where the fault changes from placing basement-on-basement to basement-on-sediment in the near-surface. The fault dip should continue to shallow with progressive overthrusting into these unconsolidated sediments. Consequently, the thickness of footwall sediments will have a large control on the width of the fault wedge in the hangingwall at scales ≤10³ m. This is observed for the Waikukupa Thrust of the Alpine Fault ~4 km southwest of the study area where exposures document decreases in basal thrust dip from 30° at low elevations to horizontal at the highest terminus [Norris and Cooper, 1997]. Detailed field mapping by Norris and Cooper [1995] at Docherty Creek has shown that from south to north the oblique-thrust segment bends from a northeast-striking moderately-dipping orientation to a north-northeast-striking shallowly-dipping orientation, while becoming more dip-slip in nature to the north. In addition to topographic controls, I propose this greater amount of thrusting over footwall sediments on
the northern end of the oblique-thrust segments might also significantly influence fault dip. The decreasing dip of the basal thrust as it progressively overthrusts footwall sediments becomes harder to accommodate oblique motion and likely encourages a rotation in fault strike or parallel partitioning. The geometry from DFDP-1 discussed above and my observations suggest the serial partitioning described by Norris and Cooper [1995] may be rooted considerably shallower than the 4 km maximum depth they calculate as the limit of stress field perturbation due to surface topography.

3.5.3 Third-order geometry of the central Alpine Fault from the LiDAR study area

The LiDAR-interpreted fault traces provide data from which I interpret the surface rupture zone on the central Alpine Fault to be reliably documented for the first time. By combining fault traces revealed in the LiDAR imagery with field observations, I can identify a smaller scale of parallel partitioning superimposed on the second-order serial partitioning. Although Norris and Cooper [1997] defined serial and parallel partitioning, they proposed them as alternative geometries for the accommodation of near-surface oblique motion. Here I demonstrate that they coexist at different scales along the same fault zone.

I confirm that active faults within the Western Province (AUS plate) are rare in the central Alpine Fault zone. As the Alpine Fault dips shallowly to moderately southeast with depth, fault traces south and east of the Alpine Fault are rooted within the base of the PAC Plate (the main Alpine Fault plane). Thus the main trace of the Alpine Fault tends to correspond to the most basin-ward (basal) fault trace, which still accommodates the most slip at the surface, with partitioning occurring only in the hangingwall.

The dynamic landscape only records a fraction of the expected horizontal and vertical displacements expected on the Alpine Fault in this area (~27 mm/yr and ~10 mm/yr respectively). Glacial trimline and moraine crest elevations indicate extensive glaciation in this area during the Last Glacial Maximum (LGM). As ice would have completely covered the range-front here [Cox and Barrell, 2007], all fault traces are presumed to be post-LGM in age. Near Arthur Creek (Figure 3.4D), incised channel offsets associated with well-preserved fault scarps protected from the Whataroa River and glacial scouring record about half of the expected strike-slip motion since the Last Glacial Maximum; most localities are ambiguous or record significantly less displacement. Fault offsets of post-LGM surfaces require most of the accumulated displacement during this time to occur on the basin-ward fault traces, which are
especially prone to fluvial erosion and slope failures. If I characterize the central Alpine Fault as a 055°-striking 50° southeast-dipping fault with an 071°-trending slip vector and a > 8 km deep brittle-viscous transition [O’Keefe, 2008; Boese et al., 2012], then I would expect at least 22 km of displacement to have been accommodated within brittle fault rocks when they arrive at the surface. Surface exposures of the Alpine Fault thrust over sediments (e.g., Gaunt Creek, Hare Mare Creek, Stony Creek, Martyr River) commonly show discontinuous excised clay gouges and ultracataclasites adjacent to an active 1–10 cm-thick continuous gouge zone composed of recycled gouge and reworked glacial till and fluvioglacial gravels. Conversely, the fault gouges in the fault core at depth were formed as a result of hydrothermal processes [Warr and Cox, 2001] during > 17 km of displacement and are smectite-rich. These gouges are frictionally weak and therefore easily accommodate localized fault slip [Boulton et al., 2012]. However, when the fault plane begins to overthrust sediments, the weak fault gouges developed at depth during accumulated displacement are abraded along its displaced contact with footwall sediments (the incipient fault plane), and a new gouge is formed from older fault gouges and footwall sediments. These new gouges will be frictionally stronger than those inherited from depth. Consequently, I expect an increase in frictional resistance to shear on the basal thrust in the near-surface (Figure 3.9C). In particular a component of the strike-slip plate motion is unable to be accommodated on the high friction low angle surfaces and so is partitioned onto the fault traces at the rear of the wedge (e.g., non-thermally weakened model of Koons et al., 2003). Thus, deformation becomes distributed into the lower rock mass strength materials in the adjacent footwall and fault rock wedge above (which is highly-fractured due to off-fault damage developed at low confining pressures). New faults form in the hangingwall to accommodate the deformation, which are expected to be the 3 orientations of faults in Figure 3.7B. I observe relatively few fault traces of note in the strongly anisotropic mylonites and Alpine Schist which have comparably high rock mass strengths, and expect the rear of the fault wedge to generally correlate with the fault-ward transition to less anisotropic fault-damaged rocks (i.e., the damage zone). The width of the fault wedge towards the range will therefore be limited by the thickness of the damage zone.

The result is a near-surface (~300 m-scale) asymmetric positive flower structure, or fault wedge, rooted on a planar moderately southeast-dipping dextral-reverse fault plane at depth (Figure 3.9). Most fault movement at depth probably occurs on a narrow fault core 1–2 m-wide and it is only in the upper ~300 m that partitioning into the transpressional flower structures occurs. This fault wedge is bounded by a basal dextral-thrust that tends to shallow as it overthrusts unconsolidated sediments in the footwall and dextral-normal faults that
bound its rear. Within the fault wedge, dextral-thrust faults concentrate along the front of the fault wedge and dextral-normal faults along the rear, consistent with extrusion of the fault wedge (Figure 3.9). The fault wedge itself consists of incohesive sediments and semi-cohesive fault rocks. I propose the width of the fault wedge is roughly controlled by the depth to basement in the footwall and the width of fault-damaged rock in the hangingwall.

Figure 3.9 Geometric models. (A) Oblique three-dimensional model based on geometry near Franz Josef (Figure 3.4C). Topographic profile is derived from the LiDAR DEM. The relative sizes of the arrows show how slip is partitioned relative to a stationary AUS plate. Lines on fault planes show expected slip vectors from field data and calculations. Plan views show expected slip vector partitioning on the fault wedge boundaries (does not account for deformation within the fault wedge). (B) Fault-perpendicular cross section through front of model in (A) showing fault wedge structure. (C) Conceptual model showing relative strengths of materials. Note the significance of the depth to basement in the footwall in controlling the basal fault dip and the width of the fault damage zone in controlling the width of the fault wedge.
The fault wedge deforms by both discrete faults and distributed folds (Figure 3.9). The resolution of the LiDAR reveals that anticlinal ridges are composed of smaller scale en echelon anticlines with a left-stepping sense, opposite to what would be predicted for anticlines produced during dextral wrench tectonics [e.g., *Sylvester*, 1988], suggesting they are more closely related to transpression and fault wedge geometry. Strengthening of the basal oblique-thrust as it approaches the surface could result in a displacement gradient on the basal plane and cause some shortening to be distributed as folding between the two fault planes bounding the fault wedge. The average fold axis trend is 055–060°, slightly oblique to the overall fault wedge orientations of about 048°. A similar analog is observed on a large scale along the Hikurangi Subduction Zone on the east coast of the North Island where regional folds and strike-slip faults within the accretionary prism trend parallel to the trench to indicate trench-perpendicular shortening despite the overall oblique displacement [e.g., *Lamb and Vella*, 1987]. The anticlines in my study are not produced by classic wrench tectonics, but by the kinematics of parallel-partitioned oblique thrust wedges.

While the wedge-bounding dextral-normal faults are kinematically stable over many earthquake cycles, the oblique plate boundary movement gradually transports them up the basal oblique-thrust plane. This causes the fault wedge to narrow until new dextral-normal faults form further back. Whatever the mechanics, the basal fault is usually still oblique and accommodates the most slip [*Norris and Cooper*, 2007].

### 3.5.4 Additional third-order controls on the central Alpine Fault

It is expected the near-surface structure of the Alpine Fault will vary throughout glacial and interglacial periods. Removal of sediments due to glacial abrasion could act to reduce the width of the fault wedge, as a thinner sediment package would give less distance for the fault plane to rotate to a lower dip. The mass of extensive ice cover during a glaciation would serve to lessen the stress perturbations caused by the deep valleys cut into the hangingwall, potentially significantly enough to reduce the development of a serially-partitioned fault zone. Extensive post-glacial aggradation of river and glacial sediment would promote a wider near-surface fault zone as the fault shallowly overthrusts a thicker sediment package in the footwall and there is a greater availability of sediment to be incorporated in a “bulldozed” fault wedge. Periods of degradation should focus the width of surface ruptures by rendering oblique-thrust wedges subcritical, as previously discussed by *Norris and Cooper* [1995, 1997].
At first glance, the predominance of range-ward facing dextral-normal scarps on a transpressive plate boundary may seem counterintuitive. Late stage uphill-facing normal fault scarps are also ubiquitous in the Charwell region of the Hope Fault in the northern South Island, despite the fact that fault accommodates predominantly strike-slip-transpressive motion [Eusden et al., 2005]. Eusden et al. [2005] proposed a model in which a 1 km-wide fault wedge is extruded between thrust and normal faults, then collapses to produce a secondary wedge comprising late normal faults in a 2 km-wide wedge in an unsupported hangingwall. In contrast to observations on the Hope Fault, I find that no obvious cross-cutting relationships exist between the observed fault traces to suggest the near-surface fault zone has evolved (i.e., partitioning has evolved with time). I suggest normal-motion traces observed on the central Alpine Fault arise purely geometrically to accommodate a more-or-less stable extruding fault wedge comprising incohesive sediment and semi-cohesive fault rocks. These traces are thus coseismic and not postseismic collapse features as on the Hope Fault. The predominantly strike-slip nature of many of the apparently “normal” traces here also argues against a collapse mechanism. The fact that some of these fault scarps have faces greater than 10–15 m-high, and a characteristic earthquake here generates < 1.5 m-throw, suggests that once established, reactivation of these faults is preferred over formation of new faults (as also demonstrated independently in the present study’s slip tendency analysis). This suggests that the fault configuration has been more-or-less stable since the LGM. I predict future earthquake surface ruptures will appear on or near known fault traces.

Second-order and third-order structures are dependent on the first-order controls of geometry, continuity, and plate motion kinematics of the fault at depth. The third-order structure of a given area is strongly influenced by the position on a second-order serially-partitioned segment as the orientation of the basal thrust or strike-slip segment will have an influence on the geometry of the near-surface fault wedge. Third-order transpressional flower structures are most obvious on oblique-thrust segments (e.g., McCulloughs Creek, Darnley Creek, Franz Josef), while strike-slip segments host fault wedges bound by parallel strike-slip faults (e.g., near Arthur Creek and upper Waitangi-taona River). In this sense there is a hierarchy of influences on fault segmentation from continuous fault planes and shear zones at depth (< $10^6$–$10^4$ m-length scales) to zones of partitioned faults in the near-surface ($10^3$–$10^0$ m-length scales).

The near-surface complexity in this region has formed despite paleoseismic evidence that the Alpine Fault ruptures during one through-going ~M 7–8 event every c. 100–500 years [Yetton,
The complexity therefore illustrates that during these large to great oblique-motion fault ruptures rooted in basement, slip does not remain localized when the fault propagates through unconsolidated surficial sediment and fault-damaged rock under low confining pressures. Near-surface complexity in the upper 1 km of the crust here is insufficient to halt earthquake rupture propagation (as proposed at different scales by Norris and Cooper, 1995; Sutherland et al., 2007a).

### 3.6 Conclusion

I use LiDAR data to contribute to a structural hierarchy of along-strike observations on the Alpine Fault.

At the first-order (< $10^6$–$10^4$ m) the Alpine Fault is a kinematically-driven system in which seismic loading and thermal weakening by advection and exhumation cause localization of oblique slip on a single relatively-planar, moderately to steeply southeast-dipping shear zone at depth [Koons et al., 2003; Ellis et al., 2006]. This orientation is inherited above the brittle-viscous transition due to the anisotropic fabric developed at depth. Development of weak clay minerals and associated high-fluid pressures contribute to a localized fault zone in the upper crust [Warr and Cox, 2001]. Consequently, at first-order the structure has a remarkably straight and continuous surface expression with no step-overs greater than 1 km [Sutherland et al., 2007a].

At the second-order ($10^4$–$10^3$ m) the central section of the Alpine Fault (where uplift rates are high) is partitioned along-strike into sequenced (i.e., serial) oblique-thrust and strike-slip segments that arise due to stress perturbations from hangingwall topographic variations in a transpressional regime on a dipping structure [Norris and Cooper, 1995, 1997]. Imbricate thrusts at this scale form as a result of rapid river incision reducing the wedge taper below its critical value [Norris and Cooper, 1997] and/or decreasing dip of the basal thrust as it progressively overthrusts footwall sediments.

At the third-order ($10^3$–$10^0$ m) asymmetric positive flower structures in the hangingwall are bound by parallel-partitioned dextral-normal and thrust faults rooted at shallow depths (< 600 m) on a planar, moderately southeast-dipping, dextral-reverse fault plane. These parallel-partitioned fault wedges form with characteristic widths of 200–600 m bounded by and containing kinematically stable fault traces. Likely influences on third-order structure include thickness of footwall sediments, width of the fault damage zone, increases in friction on the
basal fault with increasing overthrusting (due to abrasion of fault rocks), a decrease in basal fault dip in the near-surface, sea level, glacial cycles of sedimentation, abrasion and ice cover, and changes in stress orientations towards a free surface.

A fault kinematic analysis can predict the orientations of fault traces observed and indicate topography is not a significant factor at this scale. Most fault traces observed have large offsets indicating they are the product of many surface-rupturing earthquakes. A slip tendency analysis confirms the fault trace orientations are stable and are expected to accommodate repeated slip. Despite the multiplicity of near-surface traces, most displacement still occurs on a basal slip surface that preserves the hangingwall fault rock sequence. I predict future earthquake surface ruptures will appear on or near fault traces identified in this study.
Chapter 4  The Cascade Rock Avalanche: Implications of a Very Large Alpine Fault-triggered Failure, South Westland, New Zealand

Figure 4.0  Oblique aerial photograph looking southeast across the Cascade River (foreground) towards the Southern Alps. Notice the 150 m high partially vegetated arcuate scarp at center left defining the ~0.5 km$^3$ Mt Raddle deep-seated bedrock slope failure in Dun Mountain Ultramafic Group peridotite, which dominates the lower half of Mt Raddle (ultramafic peak at center left). The nearby 0.75 km$^3$ Cascade rock avalanche (off photo to the left) indicates such deep-seated bedrock failures can be precursors of very large catastrophic collapse, and should be considered a significant hazard in the Southern Alps. Haast Schist of the Olivine Range in the distance. Mt Aspiring (3033 m) forms the prominent peak on the skyline. Photo taken by L. Homer (CN5055B). Photo courtesy of GNS Science.

Abstract

Catastrophic deep-seated rock slope failures (e.g., rock avalanches) can be particularly useful proxies for fault rupture and strong ground motion, and currently represent an underappreciated seismic hazard of earthquakes in New Zealand. This study presents observations of the previously undescribed Cascade rock avalanche (CRA), a ~0.75 km$^3$ single-event, long-runout, catastrophic failure interpreted to have been coseismically triggered by a large to great earthquake at c. 660 AD on the Alpine Fault. Despite its size and remarkable preservation, the CRA deposit has been previously identified as a terminal moraine and fault-damaged outcrop, highlighting the common misinterpretation of similar rock avalanche deposits. Comparisons are drawn between the CRA and other Alpine Fault-attributed rock avalanches, such as the better-studied c. 860 AD Round Top rock avalanche, to re-assess coseismic rock avalanche hazard. Structural relationships indicate the rock mass comprising the CRA may have formerly been a portion of a larger (~3 km$^3$) rock slope failure (RSF), before its catastrophic failure along a deep-seated gravitational collapse structure (sackung). Sackungen and RSFs are common throughout the Southern Alps and other mountainous regions worldwide; in many cases they should be considered potential precursors to catastrophic failure events. Two masses of rock in the Cascade River valley show precursory signs of potential catastrophic failures of up to ~2 km$^3$; a similar mass may threaten the town of Franz Josef.

4.1 Introduction

The South Island of New Zealand lies astride an active obliquely-convergent plate boundary associated with a steep glaciated mountain belt comprising the Southern Alps and Fiordland. In this dynamic landscape, complex feedbacks operate among geological processes that create relief (e.g., tectonic deformation, isostasy), and erosional and depositional processes that reduce relief (e.g., fluvial incision, mass wasting, glacial erosion) [e.g., Koons, 1989; Norris et al., 1990; Figure 1.3 and 1.4]. A particularly dramatic example of this is the 1991 Mt Cook rock avalanche in which the summit of New Zealand’s highest mountain was instantly lowered by 10 m [McSaveney, 2002]. Landslides and other rock slope failures (RSFs) can be triggered by a variety of mechanisms including heavy rain [DeVita et al., 1998; Glade, 1998; Crozier, 2005], earthquakes [Keefer, 2002; Hancox and Perrin, 2009], freeze-thaw cycles and glacial recession [Matuoka and Murton, 2008; Allen et al., 2010; McColl, 2012], all of which have been identified as major causes in New
Zealand. RSFs can also occur without an obvious trigger, as evidenced by the 1991 Mt Cook rock avalanche [McSaveney, 2002], 1999 Mt Adams rock avalanche (190 km northeast of the study area and 4 km southeast of the Alpine Fault) [Hancox et al., 2005], 2007 Young River landslide (50 km east of the study area) [Bryant et al., 2010; Massey et al., 2011] and 2008 Vampire rock avalanches (near Mt Cook) [Cox and Allen, 2009]. However, in seismically-active regions with mountainous landscapes, strong earthquake shaking is the most likely trigger for very large, catastrophic RSFs (e.g., rock avalanches) [Keefer, 2002; Strom and Korup, 2006; Hancox and Perrin, 2009].

Major earthquakes (M > 7.0) in steep-sloped mountainous landscapes can generate strong ground accelerations, which can trigger proximal deep-seated RSFs by creating deep-seated failure surfaces or exceeding the shear resistance on existing ones [e.g., Keefer, 1984, 1994, 2002]. Additionally, tectonic damage of bedrock associated with a fault trace can facilitate large rock slope failure proximal to the surface trace of a major earthquake-producing fault [Korup, 2004; Brideau et al., 2009]. Landslide spatial densities associated with earthquakes have been shown to scale linearly with peak ground acceleration and decrease with distance from the earthquake epicenter [Menier et al., 2007]. In New Zealand, the historical coseismic landslide catalogue shows that landslides with volumes greater than $10^8$ km$^3$ occurred only in M > 7.5 earthquakes with Modified Mercalli Intensity (MMI) greater than or equal to IX [Hancox et al., 1997, 2002]. The present study focuses primarily on very large, catastrophic, deep-seated rock avalanches on the western rangefront of the Southern Alps, South Island, New Zealand, many of which have resulted from failure of glacially-steepened, tectonically-damaged bedrock triggered by ground motion generated by Alpine Fault ruptures. These rock avalanches can create dammed lakes which can fail catastrophically, and have long-runouts (~5 km) and large affected areas (~10 km$^2$). The aim of investigating these earthquake-triggered rock avalanches is to improve the paleoseismicity dataset, determine the principal factors that have resulted in failure of these rock avalanches, and help identify locations where future failures are likely, with the aspiration of mitigating the hazards posed by a major Alpine Fault earthquake.

4.1.1 The Alpine Fault – a potential driver of slope failures

The Alpine Fault is the 880 km long (~600 km onshore) major active transpressive plate boundary structure spanning the length of the South Island of New Zealand (Figure 4.1A). It
Figure 4.1  Tectonic and Geomorphic Setting. (A) Major plate boundary structures through the South Island of New Zealand. Plate motion vectors in mm/yr from NUVEL-1A [DeMets et al., 1994]. Hillshade from Land Information New Zealand (LINZ) data. Abbreviations used: Australian (AUS), Pacific (PAC), Subduction Zone (S.Z.), Round Top rock avalanche (RT), Mt Wilberg rock avalanche (MW), Franz Josef (FJ), Mt Cook (MT), Southern Alps (SA), Young River landslide (YR), John O’Groats rock avalanche (JO), Fiordland (F), Green Lake landslide (GL). (B) Alpine Fault-related structures and slope failures in the middle reaches of the Cascade River Valley superimposed on a hillshade derived from 15 m DEM created by the University of Otago – National School of Surveying utilizing LINZ data. Hope Blue River Range rock avalanche and upstream inactive gully/slip systems after Karp [2004].

Currently accommodates ~75% of total relative motion between the Pacific and Australian plates (primarily strike-slip motion) [Berrymam et al., 1992; Norris and Cooper, 2001; Sutherland et al., 2006; Barnes, 2009; Langridge et al., 2010]. Along much of its onshore length, the Alpine Fault delimits the western edge of the steep rangefront of the > 600 km long Southern Alps/Fiordland mountain belt. The fault is thought to rupture in large to great (≤ Mw 8) earthquakes with a mean long-term recurrence interval of c. 350 years, and last ruptured in AD 1717 [Yetton, 1998; Yetton et al., 1998; Wells et al., 1999; Sutherland et al., 2007a; Berrymam et al., 2012a]. The 30% probability of a ~Mw 8 Alpine Fault earthquake in the next 50 years
[Sutherland et al., 2007a; Berryman et al., 2012a] poses a significant national hazard to New Zealand; better understanding of past Alpine Fault-triggered slope failures and constraints on possible precursory indications of future events are sorely needed.

While many of the effects of an Alpine Fault earthquake will be instantaneous and catastrophic, some effects will persist for many years after. Alpine Fault-triggered slope failures may directly damage West Coast townships’ lifeline infrastructures, such as highways, bridges, housing and irrigation networks, and potentially isolate settlements. Landslide-dammed watercourses may also threaten those who live downstream, as well extensive meter-scale aggradation due to rivers responding to major sediment inputs [Robinson and Davies, 2013]. The post-seismic landscape response (e.g., flux of earthquake-generated sediment) has been shown to persist for c. 50 years after an Alpine Fault earthquake [Howarth et al., 2012; Robinson and Davies, 2013]. Damage to infrastructures (roads, bridges, buildings, trails) and scenic value (visible landslide scars, landslide debris, reduction in water clarity) will have a particularly significant effect on the livelihood of West Coast society, which presently has a growing tourism industry heavily reliant on nature-based tourism [Orchiston, 2012]. The lack of a major historical earthquake on the Alpine Fault means there is large uncertainty in the coseismic ground accelerations and resulting impact on West Coast society that can be expected during a rupture. Proxies for strong ground motion and geomorphic damage, such as the size and distribution of large RSFs triggered by previous Alpine Fault earthquakes, are one way to help reduce this uncertainty.

### 4.1.2 Slope failures in the Southern Alps

Slope failures play a significant role in the erosion of New Zealand's Southern Alps [Korup et al., 2004; Horius et al., 1997]. Extensive progradational coastal dune complexes occur from the Hollyford River (50 km southwest of the study area) to the Waitaha River (210 km northeast of the study area) and have been related to large-volume river-transported sediment pulses from the Southern Alps following major Alpine Fault earthquakes [Wells and Goff, 2006, 2007]. Alpine Fault earthquakes have been responsible for ~27% of the sediment flux from the Lake Paringa catchment (sourced on the western rangefront of the Southern Alps, 85 km northeast of the Cascade River) over the past 1100 years [Howarth et al., 2012], indicating that Alpine Fault earthquakes are a major driver of landscape evolution on the western Southern Alps, despite high rainfalls and high modern (i.e., late in the seismic cycle) sediment loads in rivers [Hicks et al., 1996]. Abundant large (> 1 km²) landslides in the Southern Alps and
Fiordland are often attributed to Alpine Fault earthquake events [Bull, 1996; Yetton, 1998; Hancox et al., 2002; Hancox and Perrin, 2009] and are likely to comprise a significant component of the post-seismic sediment flux [Korup et al., 2004; Davies and Korup, 2007]. In reality, some of these large landslides are likely to be the result of other significant earthquake sources in the area, such as the Puysegur subduction zone [e.g., Hancox et al., 2003] and faults in the Southern Alps [Cox et al., 2012a].

The most complete inventory (n = 778) of large (> 1 km$^2$) landslides in the Southern Alps/Fiordland region to date is by Korup [2004, 2005a]. Hancox and Perrin [2009] focused on 39 rock slides $> 10^7$ m$^3$ in Fiordland, including 12 slides $> 10^8$ m$^3$, and the 27$x10^9$ m$^3$ Green Lake landslide. Although the latter was certainly primed by the presence of a pre-existing fault zone and glacial processes, Hancox and Perrin [2009] used a stability analysis to argue for a seismic trigger (and against a groundwater level trigger); they concluded an Alpine Fault earthquake was the most likely trigger mechanism, despite that this landslide is a distance of 80 km from the fault. The 55$x10^6$ m$^3$ Falling Mountain rock avalanche is a well-documented event associated with the 1929 M 6.9 Arthurs Pass earthquake [McSaveney et al., 2000]. The pre-historic Round Top (RT), Mt Wilberg, Hope Blue River Range (HBRR) and John O’Groats rock avalanches (all $> 40x10^6$ m$^3$ and within 3 km of the Alpine Fault) have all been attributed to major earthquake events on the Alpine Fault [Wright, 1998b; Korup, 2004; Chevalier et al., 2009; Hancox and Perrin, 2009; Dufresne et al., 2010].

### 4.1.3 Slope failures in the Cascade valley

In its middle reaches, the Cascade valley follows the Alpine Fault for a distance of ~11 km along which slope failures are abundant (Figure 4.1B). Near to where the Alpine Fault and related faults enter the Cascade valley from the Duncan valley to the southwest, Korup [2004] identified two largely-vegetated (inactive) complex gully/slip systems (Saddle Creek and the next tributary 2.5 km to the south). He correlates their location here to fault zone weakening of the metasedimentary rock units. A ~0.5 km$^3$ coherent mass of rock west of Mt Raddle has slid over 150 m vertically and appears to be still active or dormant based on the good preservation of its scarp (shown in green on Figure 4.1B; first identified by Korup, 2005a).

Southwest of Mt Delta, Korup [2004] identified the very large HBRR rock avalanche formed in Greenland Group metasediments northwest of the Alpine Fault (shown on Figure 4.1B). The headscarp of this rock avalanche is a steep arcuate bowl some 0.5 km high and 1.5 km across. Korup [2004] estimated a deposit area of ~9 km$^2$, but the extent of the deposit is hard
to constrain due to the subdued topography, thick vegetation, and overlapping glacial moraine, river and alluvial fan deposits. Based on limited fieldwork and aerial photo interpretation it is suggested the area of this deposit is probably closer to 6 km$^2$, similar to the extent shown in Rattenbury et al. [2010] at this writer’s suggestion. The total area of the middle Cascade valley covered by slope failures is $\sim$20 km$^2$. Although more difficult to constrain, the total volume of material shed by these slope failures may be in excess of 2 km$^3$.

This study presents observations on the previously undescribed Cascade rock avalanche (CRA), a well-preserved $\sim$0.75 km$^3$ single-event, long-runout, catastrophic failure in the Cascade valley in South Westland, New Zealand. Useful comparisons are drawn between the CRA and other NZ rock avalanches, particularly the well-studied 0.04 km$^3$ RT rock avalanche [Wright, 1998b; Dufresne et al., 2010] on the Alpine Fault 250 km northeast of the Cascade River. Evidence from fault trenches and tree-ring chronologies suggest that at least 375 km strike-length of the Alpine Fault ruptured in the AD 1717 earthquake [Wells et al., 1999]. However, the along-strike extent of previous surface rupturing events is not as well known in the current paleoseismicity catalogue. The present study argues for a coseismic trigger of the Cascade and RT rock avalanches, the ages of which then contribute to the Alpine Fault paleoseismicity catalogue.

### 4.2 The Cascade rock avalanche

The large region of hummocks and undrained ponds comprising the CRA deposit has long been noted as a significant geomorphological feature of the Cascade valley [e.g., Turner, 1930]. The bulk of the rock avalanche deposit was first interpreted as a terminal moraine by Turner [1930], who proposed it would have to be a considerably younger feature than the glacial deposits at the Cascade Plateau near the coast. In a botanical study on the middle Cascade River valley, Lee et al. [1983] observed that the hummocks contained material primarily of ultramafic origin and interpreted the area to be composed of a mixture of morainic till and rockfall debris dating to the Otiran Advance of the last glaciation. Sutherland [1995] and Sutherland et al. [1995] echoed Turner’s terminal moraine interpretation, attributing a latest Last Glacial Maximum (LGM) age for the deposit. Campbell [2005] invoked thrust faults to interpret the region as outcrops of Pacific plate fault-damaged basement overthrust onto the Australian plate. The present study utilizes field observations and interpretation of stereoscopically-paired aerial photography to identify the region as part of a 9.1 km$^2$ rock avalanche deposit shed from the slopes of Martyr Spur (Figure 4.2).
4.2.1 Deposit morphology

Though thickly vegetated, the surface morphology of the CRA deposit is well-preserved and is clearly indicative of a single event, as particularly evidenced by the even pattern of radiating longitudinal ridges (Figure 4.3). Hummocks, small ponds and longitudinal ridges are the most striking morphological features present.

Over 100 hummocks have been identified from stereoscopic aerial photography. Hummocks decrease in size and height away from the source scarp. The highest single hummock (138 m
elevation) has a minimum relief of ~50 m above the surrounding portion of the deposit. The most distal hummocks tend to be less elongate and lack a defined ridge. In the middle of the deposit, many of the hummocks are grouped into clusters that form striking longitudinal ridges with a radial pattern of their long axes away from the source scarp (Figure 4.3). These radiating longitudinal ridges have a spreading angle of 75°. By tracing the ridge axes back to a common point it is found that those around Smiths Ponds have an origin slightly distinct from that of the longitudinal ridges on the fringes of the radial pattern (Figure 4.3). The common origin of the Smiths Ponds radiating longitudinal ridges is at a position close to the base of the source scarp, while the major origin of the enveloping ridges is closer to the center of gravity of the CRA mass before failure (~1 km to the southeast). By fitting a line to the two origin points, a fall line of 315° can be estimated. This is very close to perpendicular to the 055° strike of the Alpine Fault here. Similar patterns of longitudinal ridges and hummock distribution have been noted for the RT rock avalanche 250 km to the northeast [Dufresne et al., 2010] and The Hillocks rock avalanche, 70 km to the south of the study area [McColl and Davies, 2011]. Lakes and ponds on the rock avalanche deposit are characteristically dentate, reflecting the shapes of the bounding hummocks. The shores of these water bodies consist of subangular to subrounded pebbles, cobbles and occasionally boulders, all locally derived from the surrounding hummocks, which allows the distribution of rock types within the rock avalanche to be mapped. Many of these water bodies appear to lack outlets and water levels fluctuate depending on rainfall. Most of these ponds are 1-5 m-deep when full. The deepest portions of the water bodies usually contain fine silt.

The trace of the Alpine Fault is well-defined in aerial imagery both northeast and southwest of the rock avalanche, but much less so within the deposit, likely reflecting the deposit’s young age. A lack of significant offset where the headscarp crosses the fault also suggests a youthful age. A few small ponds occur near where the Alpine Fault cuts across the head of the deposit, and may be ponded against a fault scarp.

The rock avalanche deposit has an area of ~9.1 km² and an estimated volume of ~0.75 km³ (calculated assuming a simple flat basal surface of the CRA deposit at ~0 m above sea level). Fluvial erosion and modification by the Cascade River, particularly along the northern edge of the deposit has somewhat lessened the observable volume (and potentially runout distance) of the deposit. Fluvial modification likely explains the lack of hummocks in the northern portion of the deposit.
The surface of the rock avalanche supports a ~12 m-tall mixed podocarp-beech forest [Lee et al., 1983]. Lee et al. [1983] noted that the canopy height is lower on hilltops and varies from one hill to another, which the present study suggests is likely a function of the underlying
lithologies and consequent variations in soil development rate. The only portions of the rock avalanche deposit that are sparsely vegetated are the shorelines of the ponds, the steep cliffs along the Cascade River, and an active 150 m-high slip near the prominent bend in the Cascade River within the deposit. Exposures in these regions allow useful insights into the internal structure of the rock avalanche.

4.2.2 Dammed lake

The present morphology suggests that at the time of deposition the lowest point across the rock avalanche would have been at its west margin at an elevation of ~80 m, or at least 50 m above the present river level. At such an elevation, the rock avalanche would have dammed the Cascade River to at least the bottom of the gorge through the HBRR rock avalanche 5 km upstream to create a 4.6 km² lake (Figure 4.1B), (but possibly much further if there has been significant post-rock avalanche aggradation). There is no evidence for a former outlet along the western margin, but it is possible one could be buried under post-rock avalanche alluvium and rockfalls shedding from the Hope Blue River Range. The Cascade River has cut a 5 km-long steeply-walled gorge through the middle of the rock avalanche deposit. The CRA gorge has a characteristic width of ~100 m and a maximum height of ~60 m. In the lower reaches of the gorge 3 m of river gravels overlie a strath surface cut into the rock avalanche deposit ~8 m above river level. The fluvially-modified portion of the CRA deposit may be related to a former outflow of the CRA gorge at the elevation of this strath surface before the river migrated and incised further east (Figure 4.3).

4.2.3 Runout

The horizontal distance from the head of the scarp to the toe of the landslide ($L_{\text{max}}$), is 4800 m, with a vertical fall of 620 m from the top of the headscarp to the lowest point at the toe of the deposit ($H_{\text{max}}$), which yields a $H_{\text{max}} / L_{\text{max}}$ value of 0.129 (terminology after Legros, 2002). These parameters all plot within a range of $L_{\text{max}}$, $H_{\text{max}}$, and V (volume) empirical relations compiled for subaerial non-volcanic landslides (see Figure 3 of Legros [2002]. The runout distance from the bottom of the source slope (denoted $L^*$ here) is 3.5 km. The average slope of the deposit across this runout distance ($L^*$) is about 0.06 or 4°. The deposit reaches a maximum thickness of at least 220 m. The distance from the pre-failure center of gravity to post-failure center of gravity for the slide mass is ~1400 km ($L$), with an associated vertical fall of ~180 m ($H$). The latter measurements can be used to calculate an apparent coefficient of friction ($H/L$) for the basal sliding surface of ~0.129 (Table 4.1). For comparison, Wright
[1998b] determined an H/L value of ~0.24 for the main RT rock avalanche deposit. Many CRA and RT rock avalanche statistics including runout length, drop height, substrate slope, fall line and spreading angle are comparable, reflecting their similar structural and geomorphic settings (Table 4.1).

The consistent radial pattern of the longitudinal ridges and evenly distributed hummocks of the rock avalanche deposit (e.g., Figure 4.2 and 4.3) suggest the runout surface was essentially flat with no major obstructions (e.g., glacial ice, moraine, or glaciated bedrock knolls such as Charlie's Bump) in its path; this pattern also indicates the CRA occurred as a single event. It is assumed that the runout surface was dominated by low-lying beech-covered river floodplains or mossy swamps typical of adjacent areas of the lower Cascade valley. There is a defined break in slope where hummocks of the rock avalanche deposit come in contact with the steep bedrock hillslopes of Mt Delta, suggesting there might not have been significant run-up of

| Table 4.1 Cascade and Round Top Rock Avalanche Statistics. Key statistics of the Cascade and Round Top rock avalanches. Abbreviations and definitions: A (area), V (volume), L<sub>max</sub> (map distance from the headscarp to distal edge of rock avalanche deposit), L* (map distance from the base of the source slope to distal edge of rock avalanche deposit), L (map distance from the pre-failure center of gravity to center of gravity of post-failure deposit), H<sub>max</sub> (vertical distance between the highest elevation at the source to lowest elevation of the rock avalanche deposit), H (vertical difference between the pre-failure center of gravity and the center of gravity of post-failure deposit). Some Round Top rock avalanche statistics after Dufresne et al. [2010] and Wright [1998b]. |
|---|---|
| | Cascade | Round Top |
| A (km<sup>2</sup>) | 9.1 | 5.6 |
| V (10<sup>6</sup> m<sup>3</sup>) | 750 | 45 |
| Age (2σ cal AD) | >540 (+104/-118) † | 860 (+119/-95) ‡ |
| L<sub>max</sub> (m) | 4800 | 4800 |
| L* (m) | 3500 | 3500 |
| L (m) | 1400 | 1450 |
| H<sub>max</sub> (m) | 620 | 570 |
| H (m) | 180 | 310 |
| H<sub>max</sub> / L<sub>max</sub> | 0.129 | 0.119 |
| H / L | 0.129 | 0.214 |
| Substrate slope (°) | <1 | 1 – 2 |
| Average slope (% / °) | 5.1 / 2.9 | 3.3 / 1.9 |
| Fall line (°) | 315 | 334 |
| Spreading Angle (°) | 75 | 90 |

† Calibrated radiocarbon age of entrained log. A “<” sign is used because the exact age of the log at the time of entrainment is unknown. Factoring in a 120 yr estimate for the age of the log at time of entrainment yields a c. 660 AD for the Cascade rock avalanche

‡ Calibrated radiocarbon age incorporating a 95 ± 5 yr age of the log at the time of entrainment after Wright, 1998b
the rock avalanche up the valley side opposite its source area. The formerly-glaciated valley side adjacent to the rock avalanche has an average slope of 28°, which is expected to be close to the pre-failure average slope angle of the hillside at the rock avalanche.

### 4.2.4 Internal structure of deposit

The CRA deposit is sourced in three distinct lithologic units: (1) Greenland Group quartzose metasandstone and metamudstone with quartz-feldspar leucosomes and occasional granitoid intrusions, locally mylonitized within 150 m of the Alpine Fault; (2) Brook Street Volcanic Group-derived chlorite-epidote-quartz metavolcanicalastic mylonites; and (3) Dun Mountain Ultramafic Group serpenatinized peridotites and serpentinites with cross-cutting serpentine shear zones (Figure 4.3; Figure 4.4). Good outcrops at the Martyr River and Martyr Spur to the northeast of the rock avalanche and in creeks to the southwest are unaffected by rock avalanche deformation. The orientations and textures of the lithologic units adjacent to the rock avalanche can be used as proxies to infer fractures, shears, and block rotations formed by the rock avalanche during transport (Figure 4.5).

The lithologies affected by the Cascade, RT, HBRR, and John O’Groats rock avalanches are different. The RT rock avalanche involved only schist-derived mylonites and cataclasite, the HBRR rock avalanche sources only quartozfeldspatic metasandstone and metamudstone; and the John O’Groats rock avalanche 70 km southeast sources only mylonitic orthogneiss. These observations suggest that lithology and inherited rock fabric (apart from fault damage) do not play a significant role in predicting large rock avalanche occurrences in this environment.

![Figure 4.4](image) Cross section of the Cascade rock avalanche along A-A’ (see Figure 4.3) constrained by topography and field observations.
4.2.4.1 Rock textures

In this section, outcrops outside the CRA deposit and outcrops within the CRA deposit are compared. In the vicinity of the CRA headscarp (outside of the CRA deposit), the Dun Mountain Ultramafic Group rocks show a high degree of serpentinization with ubiquitous serpentine (antigorite) coating shears and curviplanar fracture surfaces. Steeply-dipping, anastomosing zones of sheared serpentine (antigorite) 1–10 m-wide surround black serpentinite (lizardite replacing dunite and wehrlite) blocks 1–30 m-across to define a large scale broken formation fabric. Pervasive anastomosing shear fabric, crude layering and blocky fractures within lozenges are typical of these outcrops and define the structure the rock avalanche inherits (Figure 4.5A).

Exposures of serpentinites within the CRA deposit have a highly brecciated block-in-matrix appearance when viewed up close, but at an outcrop-scale original layering and occasional dikes can be traced for several meters (Figure 4.5B). Despite the relatively high fracture density, entire outcrops are essentially single blocks of coherent, but pulverized, rock (i.e., the texture is not cataclastic wherein it is expected that particles are rotated and translated with respect to one another). At the hand sample-scale, jigsaw textures [e.g., Shreve, 1968] are common in more coherent portions of the unit with gaps between coherent clasts containing sand-sized (and smaller) particles (Figure 4.5C). Many of the fractures not coated in antigorite or mineralized with quartz or calcite are presumed to have formed during the rock avalanche event. Outcrops through serpentine-rich zones of shear qualitatively have a different macroscopic particle size distribution, as the fracture pattern is partially dictated by the shape of preexisting serpentinite lozenges and original shear fabric (cf. Figure 4.5C–D). Whereas the shear fabric in the serpentinites dips steeply outside the CRA deposit, within the deposit the fabric typically dips shallowly to moderately to the southeast or northwest.

Figure 4.5 Rock Textures. (A) Typical serpentinite outcrop adjacent to the Cascade rock avalanche (from the Martyr River). Notice anastomosing shear fabric, crude layering and blocky fractures. (B) Large outcrop of serpentinite within the Cascade rock avalanche. Despite the highly brecciated block-in-matrix appearance up close, original layering is continuous over several meters indicating the outcrop is highly fractured, but not cataclastic. Trekking pole is 1.1 m high. (C) Close-up of outcrop in B showing typical jigsaw texture. Gaps between coherent clasts contain sand-sized (and smaller) particles. (D) Serpentinite outcrop within the Cascade rock avalanche. Note fracture pattern is partially dictated by the shape of serpentinite lozenges and original shear fabric (sub-horizontal in this view). (E) Typical metavolcanoclastic mylonite outcrop adjacent to the Cascade rock avalanche (from the Martyr River). Notice spacing of foliation perpendicular fractures and scarcity of foliation parallel fractures. (F) Outcrop of mylonite within the Cascade rock avalanche. Note spacing of fractures and selective pulverization of epidote-rich layers (light green) over the chlorite-rich layers (dark green-black). Original mylonitic fabric is largely intact despite intense fracturing.
Outside of the CRA deposit, chlorite-epidote-quartz mylonites with a metavolcaniclastic protolith occur in a ~400 m section in the Martyr River between serpentinites and the Alpine Fault. These mylonites are locally folded, but generally the mylonitic fabric dips about 40° to the southeast (e.g., in the creek immediately south of the CRA deposit). Typical outcrops are characterized by 1–10 cm-thick alternating layers of dark green chlorite and light green epidote; foliation-perpendicular fractures are predominant, but with occasional fractures on the foliation planes themselves (Figure 4.5E). Within the CRA deposit, the mylonitic rocks have very little disruption to the original fabric, but have a higher density of fractures; epidote-rich layers are selectively fractured to a greater extent (Figure 4.5F). Mylonitic foliation within the CRA deposit dips about 50° to the southeast. Within 80 m of the Alpine Fault (outside the CRA deposit), the metavolcaniclastic mylonites are intensely fractured (cm-scale fracture spacing) and have a pervasive chlorite hydrothermal alteration; these rocks are especially prone to slope instability as evidenced by a 15 m-high cliff at the Martyr River that retreated over a meter from 2009 to 2011. Comparable rocks are not exposed in the CRA deposit.

There are no exposures through hummocks composed of Greenland Group lithologies. The Greenland Group is only seen on the surface of the rock avalanche deposit as isolated boulders, cobbles and pebbles along the shores of ponds and on top of hummocks. In situ Greenland Group rocks have a strong planar fabric that typically dips 60–90° to the northwest or southeast. The Australian plate damage zone of the Alpine Fault is ~150 m-wide and consists of sheared quartzose and pelitic mylonites that decrease in competence with distance toward the fault. Outside of the damage zone the Greenland Group lithologies are characteristically unfaulted and unfractured here. Along the north end of the river-cut gorge is an exposure of glacial till and folded silts juxtaposed against a highly fractured outcrop of sheared serpentinite; this is the only block of glacial sediments observed in the CRA deposit.

4.2.4.2 Hummock structure

Vertical 40 m-high exposures along the river gorge are laterally continuous for hundreds of meters and present some of the best cross sections through CRA deposit hummocks (Figure 4.6A). Despite the highly sheared and fractured nature, lithologic horizons are traceable for over 12 m. While it is difficult to resolve whether the serpentinite shear fabric was reactivated
Figure 4.6 Hummock Structure. (A) Vertical 40 m high exposure in the gorge cut by the Cascade River (foreground). This serpentinite outcrop is associated with a single large hummock. Discrete lithologic horizons can be traced across much of the width of the outcrop. It is unclear whether the serpentinite shear fabric was reactivated during transport of the rock avalanche deposit. The outcrop extends to river level; no basal mixed zone is exposed. (B) Chaotically mixed zone between 50 m scale block of serpentinite above and volcaniclastic mylonite below. The 4 m wide breccia zone is composed primarily of angular clasts of rotated metavolcaniclastic mylonite in a finer matrix of the same. Foliated serpentinite fabric at hammer would have easily accommodated layer-parallel slip and may have been created or reactivated during rock avalanche transport.

4.2.4.3 Transport

It is possible to track the movements of different portions of the rock avalanche mass during its transport by comparing the spatial distribution of lithologies within the CRA deposit to their source locations (Figure 4.4). Serpentinite dominates the outcrops around the river gorge and as float on proximal hummocks. Metavolcaniclastic mylonite is confined to a few
proximal outcrops near the river. Quartzopelitic Greenland Group metasedimentary rocks occur on the western margin of the CRA deposit and by volume balance must comprise a significant component of the distal portion of the deposit (Figure 4.3; Figure 4.4). Thus, the base of the former hillslope now comprises the most distal portions of the rock avalanche deposit, while the mass formerly nearest the headscarp remains proximal (Figure 4.4). This is common in rock avalanches.

Nowhere is the base or substrate of the CRA deposit exposed [cf. RT rock avalanche observations of Wright, 1998b and Dufresne et al., 2010]. In two locations rock avalanche deposit outcrops extend greater than 1 m below present river level (e.g., near outcrop in Figure 4.6A), implying that either (1) the Cascade River has not incised back to its prior base level or (2) the rock avalanche excavated its substrate during transport. Since the river level is already close to base level at this location (~15 m elevation), I suggest that the rock avalanche excavated the fluvial gravel or swampy peat substrate, analogous to observations of the RT rock avalanche deposit by Dufresne et al. [2010]. They determined the RT rock avalanche included a basal mixed zone containing entrained substrate and rock avalanche material that is greater than 3 m thick. Dufresne et al. [2010] argue that the planar surface of saturated gravel over which the RT rock avalanche slid has contributed to its final morphology and mobility (runout distance). Their arguments can be applied to the CRA by analogy. The long runout could also be explained by observations of mechanical fragmentation and comminution in the internal structure of the CRA. Davies et al. [1999] and Davies and McSaveney [2009] explain how mechanical fragmentation in granular flow at the base of large rock avalanches can exert dispersive pressures, which serve to reduce the frictional resistance on the basal surface. Although the deposit character corresponds well with that generally found in rock avalanches, the absence of an exposed basal zone in the CRA deposit prevents direct observation of intensely-fragmented material here.

In summary, transport of the rock avalanche mass produced intense jigsaw-textured fractures in large otherwise cohesive blocks which form hummocks; rock avalanche textures overprint pre-existing tectonic fractures and shears. The nature and intensity of the rock-avalanche-produced textures depend to some extent on the lithology and pre-existing structural anisotropy of the source material.
4.2.5 Deposit age

Nowhere is the sedimentary material underlying the rock avalanche exposed. My preliminary attempts to determine its age focused on searching a chaotically mixed zone between blocks near the river gorge for radiocarbon dateable material. This proved unsuccessful.

Lee et al. [1983] obtained a conventional $^{14}$C radiocarbon date of 1572 ± 60 years (NZ 4626) from a beech log “trapped in a rockfall from slopes of Mt Delta” (which would comprise only Greenland Group lithologies) that had been exposed in a 3 m high terrace cut into the edge of the rockfall (see location in Figure 4.3). They note that “the rockfall is contiguous with a large area of ultrabasic breccia,” but did not recognize the possibility that the Greenland Group lithologies could have been transported from the same source area as the ultramafic rocks (i.e., the northwest slope of Martyr Spur) instead of from Mt Delta. About 100 m to the northwest of their collection site, small rockfalls and steep creeks have deposited Greenland Group lithologies from the slopes of Mt Delta onto the CRA deposit. However, these areas are rockier with less soil development and can be easily distinguished from the Cascade rock avalanche deposit. Field examination of their collection site in this study has confirmed that the site they sampled is within the CRA, and that the deposit extends across the full length of the valley to the base of Mt Delta. This radiocarbon age has been calibrated using the CALIB 6.0 program [Stuiver and Reimer, 1993] with the Southern Hemisphere Atmosphere correction curve [Reimer et al., 2009] to obtain a median probability age of 540 AD with a lower 2σ of 423 AD and an upper 2σ of 645 AD. Unfortunately Lee et al. [1983] did not note the approximate age of the log sampled or whether the sample was collected from the sapwood or heartwood, but it is likely the sample was collected from the heartwood [W.G. Lee pers comm., 2012]. Mature beech, which would have developed such heartwood, in this environment is typically 120–160 yrs old [Wardle, 1984]; this time has to be subtracted from the total radiocarbon age to determine the time at which the tree was incorporated into the CRA deposit. Using the radiocarbon date, this study obtains a best estimate of 540 (+104 / −118) cal AD for the radiocarbon sample. Incorporating a 120 year best guess for the age of the radiocarbon sample at the time of entrainment shifts the best estimate for the CRA toward c. 660 AD.

This radiocarbon age is consistent with several observations. (1) The deposit must be younger than the youngest dated moraine deposit downriver. Sutherland et al. [2007b] obtained a $^{10}$Be exposure age of 13.8 ± 1.7 ka for a boulder atop a Last Glacial moraine deposit at 200 m
elevation on the Cascade Plateau; this is also a lower bound on the timing of this glaciation.

(2) Many trees have difficulty adapting to the relatively toxic concentrations of nickel and/or magnesium in the ultramafic soil [Robinson et al., 1996]. The establishment of forest on a surface comprising ultramafic and quartzose lithologies is expected to take longer than if no ultramafics were present. Trees on comparable ultramafic rocks in a comparable environment near Milford Sound, 75 km to the southwest were found to be 350–450 years old, suggesting a minimum length of time for mixed forest colonization on ultramafic rocks [Lee, 1992]. This then provides a minimum age for the rock avalanche deposit. (3) The Alpha Creek alluvial fan extends over 300 m onto the Cascade rock avalanche and is forested (Figure 4.3). (4) It is also noted that although its lithology is different to that of the Cascade rock avalanche, the 860 AD RT rock avalanche has comparable forest development and preservation of surface morphology. (5) The trace of the Alpine Fault is well-defined in aerial imagery northeast and southwest of the rock avalanche deposit, but is youthful where it crosses the deposit. Dextral offset of the northern edge of the rock avalanche headscarp across the Alpine Fault trace is constrained to be less than 50 m (though best guess for offset is ~30 m), which requires a maximum of 1850 years utilizing a 27 mm/yr dextral strike-slip rate for the Alpine Fault here. Further, finding a precise fall line by restoring radial longitudinal ridge axes back to a null point and comparing this to the corresponding headscarp position indicates fault offset is minimal. Together these relative dating observations independently constrain the rock avalanche event to between c. 300–2000 yrs BP, in agreement with the radiocarbon age estimation of c. 660 AD.

### 4.2.6 Source area

Comparing the distribution of lithologies within the CRA deposit to the well-mapped basement geology, and restoring the CRA deposit’s radiating longitudinal ridges back to a common point, both unequivocally pinpoint the source area. The metavolcaniclastic mylonite is a particularly diagnostic unit as it only outcrops in the Cascade River catchment at the CRA source area and for another 2 km to the southwest along the base of the Martyr Spur (Figure 4.1B). However, at these locations to the southwest there are no Greenland Group rocks outcropping at the base of the hillslope. The only location where all three major CRA deposit lithologies occur is within a deep, amphitheatre-shaped source area in the otherwise smoothly glaciated hillslope of Martyr Spur. Using a hillslope profile immediately south of the CRA source as an estimate of the pre-CRA hillslope surface, volume estimations for the CRA source area yield ~0.75 km³. This is similar to the estimate determined for the CRA deposit.
A ~25% volume increase (“bulking”) [Hungr and Evans, 2004] is expected from the initial source volume to the CRA deposit and may represent the amount of mass removed by the Cascade River in the northern portion of the CRA deposit.

The CRA has a 3 km-long asymmetrical headscarp with its long side along Martyr Spur. The average scarp height is about 250 m to the base of the slope defining the top of the rock avalanche deposit remaining on the headscarp. The headscarp is forested, indicating it has been stable for centuries. Gullies on the northern half of the scarp face indicate an inactive gully/slip system that may have opened shortly after exposure by the CRA and has since stabilized.

4.2.7 Failure mechanisms

While the trigger for the CRA is interpreted to be coseismic ground motion (see Section 4.3.1), structural priming of the mass seems to have played a principal role as a mechanism for failure. The Alpine Fault and associated fault-damaged rocks cross the lower slopes of the source area. Quasi-linear normal-sense scarps line the top of Martyr Spur with the most obvious scarp coinciding with the long axis of the CRA headscarp. This normal-sense scarp appears to be coincident with the main failure surface of the rock avalanche, so the significance of these features will now be further explored.

4.2.7.1 Sackungen

Uphill-facing scarps indicative of deep-seated gravitational slope collapse, or sackungen (singular: sackung; also known as antslope scarps, locally referred to as ridge rents), are observed worldwide in regions of high relief that have been oversteepened by past glaciations and/or uplift [e.g., Rockies: Radbruch-Hall et al., 1976; Alaska: Li et al., 2010; Himalaya: Schroder, 1998; Alps: Ambrosi and Crosta, 2006; Pyrenees: Gutiérrez-Santolalla et al., 2005; Pere, 2009]. While they are not tectonic faults, it is often proposed that seismicity can generate or reactivate sackungen [Beck, 1968; Pasuto and Soldati, 1996; Pere, 2009]. The 2002 M 7.9 Denali earthquake in Alaska produced 1–3 m of throw on sackungen; some of these movements correlated with pre-existing notches in ridges indicating reactivation [Jibson et al., 2004]. In some cases, sackungen are seen as precursory phenomena for accelerated or catastrophic failure [e.g., Kompp, 2005b]. The good exposure of these structures in alpine areas and different geometries observed in different geologic settings has led to a wide range of formation mechanisms being proposed; such as lateral spreading, flexural slip, and listric normal faulting.
[e.g., Beck, 1968; Li et al., 2010]. In New Zealand sackungen are common as ridge-parallel, 60–70° dipping ~10 m-high uphill-facing scarps, sometimes over a kilometer in length. They occur along steep ridges of schist and greywacke near the main divide of the Southern Alps and are typically paired such that the cumulative throw on uphill-facing scarps on either side of the ridge are approximately equal [Beck, 1968; Perri, 2009]. Beck [1968] observed that sackungen readily cross-cut greywacke bedding and schist foliation, and argued their orientation is not controlled by existing rock fabric.

Beck [1968] proposed three geometric models to explain sackungen formation in the Southern Alps: (1) a lateral spreading model in which sackungen rooted into the center of the ridge form a graben and the sides of the ridges spread laterally to accommodate collapse of the ridge (the ridge elevation is lowered, but the slopes of the hillsides remain unchanged; see his Figure 3A), (2) a base extrusion model in which sackungen rooted into the center of the ridge form a graben and crushed rock is extruded from the base of the ridge to accommodate collapse of the ridge (the ridge elevation is lowered and the slopes of the hillsides increase; see his Figure 3B), and (3) a listric model in which sackungen are rooted into the opposing hillside as listric failure surfaces (see his Figure 3C). His third model is the only one in which both the ridge elevation and slopes of the hillsides decrease, which allows sackungen to decrease the gravitational potential of oversteepened slopes. This study adapts a variation of his listric model in which the main failure surfaces are likely rooted into the base of the topography, to explain symmetrical sackungen distributions on ridges bounding valleys of equal elevation (Figure 4.7A).

**4.2.7.2 Asymmetrical ridges, asymmetrical sackungen**

Martyr Spur differs from the symmetrical model in that the sackungen on both sides of the ridge have west-northwest facing scarps and thus show an asymmetrical distribution. Asymmetrical sackungen have elsewhere been interpreted as resulting from previous collapse of a ridge (and thus scarps will not necessarily be symmetrical with respect to the newly-formed ridge) or from a strong anisotropy within the rock, such as bedding, foliation or unit contact [Beck, 1968; Li et al., 2010]. It is proposed that the asymmetric collapse of Maryr Spur towards the Cascade River is due to topography and not rock anisotropy. The elevation of the Martyr River opposite the rock avalanche is about 250 m higher than the Cascade River; presumably this is because glacial deepening of the Cascade valley is greater than that of the Martyr valley (a smaller tributary). Gravitational collapse is clearly favored towards the lower elevation valley; presumably this is because glacial deepening of the Cascade valley is greater
Figure 4.7  Sackungen Models. (A) Model of sackungen development on a symmetrical ridge with similar elevation valley floors on either side as common in the Southern Alps. In this model after Beck [1968], sackungen decrease the height of the ridge and lessen the steepness of the valley slopes. (B) Model proposed for the development of sackungen associated with an asymmetric ridge bounded by differing valley floor elevations. Foliation in all rock units (not shown for simplicity and clarity) is sub-parallel to the Alpine Fault. Gravitational potential prefers collapse of the ridge toward the lower elevation valley, comparable to the scenario at the Cascade and Round Top rock avalanches. In the Southern Alps, sackungen cross-cut pre-existing rock fabrics (e.g., bedding, foliation), and thus appear to have little influence on the orientation of sackungen [Beck, 1968; this study].

than that of the Martyr valley. Based on the scarp geometry, these slope collapse structures are interpreted as listric normal faults rooted into the base of the ridge near the Cascade River (Figure 4.7B).

Sackungen scarps at Martyr Spur strike north-northeast and dip ~60° west-northwest. Scarp heights vary from 2 to 100 m, with ~5 m being the norm. Where obvious, the motion is pure dip-slip. Several ponds have formed in the troughs adjacent to sackungen. The scarps reach 2 kilometers in length and typically strike parallel, but also form en-echelon or anastomosing patterns to link adjacent structures. They are spaced 50 to 250 m apart. The scarps support scrubby vegetation in contrast to the sparse vegetation cover of the adjacent ultramafic outcrop, potentially because the rock has less surface weathering on scarp surfaces or the steeper slopes have protected the vegetation from fires. Since the scarps are evenly vegetated to their bases, there is no indication of recent activity.

The RT rock avalanche similarly has the Alpine Fault at the base of its source area and, although less obvious due to its vegetated headscarp, distinct ridge-parallel sackungen on the ridge northeast and southwest of the headscarp are visible (Figure 4.8). While the dip direction of sackungen at Round Top is hard to resolve, it is noted there is a significant asymmetry to the elevations of the drainages on both sides of the Round Top ridge (~150 m
Figure 4.8  Round Top Rock Avalanche. Oblique aerial view of the Round Top rock avalanche looking roughly north. Aerial photograph is draped over a digital elevation model with no vertical exaggeration (source: Google Earth). Note comparable runout distance, deposit morphology and structural setting to the Cascade rock avalanche. Also note the presence of a sackung-like linear hillslope break associated with the headscarp region; comparable but more subtle features are present on the north side of the headscarp. Scale varies in this perspective.

on the northwest and ~380 m to the southeast). This topographic asymmetry is comparable to that across Martyr Spur, and may indicate Round Top is another scenario where asymmetric sackungen likely form (northwest-dipping only) (Figure 7B). The principal failure surface of the RT rock avalanche cuts across the strong moderately southeast-dipping foliation in the schistose mylonites here [Nathan et al., 2002].

4.2.7.3 Sackungen as precursors to catastrophic failure

Many RSFs in the region and elsewhere have been associated with pre-existing structures such as sackungen. Korup [2005b] highlighted several cases in the Southern Alps where sackungen and RSFs (his “DSGSD”) were seen as precursory phenomena for accelerated or catastrophic failure. A ~5 km² complex rock slide in the Lake Leeb and Lake Clarke region 10 km east of the CRA has a clear relationship to prominent ridge-parallel north-northwest striking, down-to-the-west sackungen in Haast Schist directly above the rock slide [Rattenbury et al., 2010]. The 2007 0.011 km³ Young River landslide, 50 km east of the study area has prominent sackungen associated with, and localized to, the headscarp area [Bryant, 2010; Massey et al., 2011; S. McColl unpublished data]. The headscarp of the 2008 Mw 7.9 Wenchuan
earthquake-triggered 0.84 km$^3$ Daguangbao landslide in China comprised a 2 km-long ridgetop linear depression (sackung) before failure [Chigira et al., 2010]. Similar pre-existing linear depressions were found at the 1999 M$_w$ 7.6 Chi-Chi earthquake-triggered 0.15 km$^3$ Tsaoling slide in Taiwan [Chigira et al., 2003]. Many of the largest coseismic slope failures have a pre-existing structural control [e.g., Keefer, 1984; Hewitt et al., 2008]. It is proposed the significance of sackungen in indicating the potential for catastrophic deep-seated slope failures may be underappreciated.

Using observations of the CRA and RT rock avalanche case studies, I now highlight three specific examples of pre-existing RSFs in the Southern Alps that may be capable of catastrophic failure.

**Martyr Spur**

There is a strong genetic relationship between the headscarp of the CRA and the largest sackung on Martyr Spur. This sackung scarp can be traced for over 2 km to the south with a throw of almost 100 m. The scarp height decreases to the south, away from the rock avalanche. The sackung intersects the CRA headscarp and seems to be congruent with the 2 km long axis of the headscarp (Figure 4.3). Slope degradation in the form of erosional gullies exists evenly along the entire length of this main sackung scarp. There are no known rock discontinuities associated with the short axis of the CRA headscarp. Adjacent to the main sackung are northeast-striking sackungen that together define a large deep-seated bedrock mass (~2 km$^3$). Together, these observations suggest the sackung formed the primary failure surface for the CRA, and that the CRA mass may have formerly been a portion of a larger (~3 km$^3$) deep-seated RSF before its subsequent catastrophic collapse. The remaining ~2 km$^3$ RSF could take part in a future catastrophic failure (Figure 4.3). Despite the potential for this remaining 2 km$^3$ mass to catastrophically fail, a considerably smaller volume event could significantly affect the Cascade River valley (Figure 4.3). Because of the barrier of the existing CRA deposit, it would only take a comparatively small volume failure (~$10^6$ m$^3$) to dam the Cascade River at the river gorge through the CRA deposit. Inundation of the Cascade River behind a potentially unstable dam would have a significant effect on settlements, fisheries and productive cattle land in the lower Cascade valley.

**Mt Raddle**

Another RSF considered a potential catastrophic failure is a ~0.5 km$^3$ mass of peridotite (herein named the Mt Raddle RSF, but previously identified by Korup, 2005a) on a major
Mt Raddle Rock Slope Failure (RSF). Aerial photograph of the Mt Raddle RSF, a large ~0.5 km$^3$ mass of peridotite that could be primed for a catastrophic failure triggered by ground motion associated with a future Alpine Fault earthquake. The RSF mass has slid down about 150m vertically; white lines denote identifiable offset of a rock unit and of a glaciated valley wall. Well-preserved lateral moraines above the headscarp indicates all slope failure across the headscarp is likely post-glacial. A catastrophic failure at this location would almost certainly dam the Cascade River. See Figure 4.1B for location.

inside bend of the Cascade valley, ~12 km southwest of the CRA (Figure 4.9). The Mt Raddle RSF mass appears relatively unserpentinized compared to the rocks immediately to the east. The mass is defined by an arcuate headscarp over 1.5 km-long. This scarp indicates the Mt Raddle RSF has thus far collapsed about 150 m vertically and horizontally toward the Cascade valley (Figure 4.9). Offset of the base of the valley hillslope across the south side of the slope failure indicates the failure surface is likely a curviplanar listric surface that is rooted at or below the elevation of the valley floor. Well-preserved lateral moraines above the headscarp and the rocky, relatively uneroded nature of the scarp indicate all slope failure
across the headscarp is probably post-glacial, and likely significantly younger. The toe of the slope is bulging and the hillslope has slip headscarps extending along its steepest slopes. Deformation of the valley floor by this RSF may explain diversion of the Cascade River from an unvegetated, recently-active floodplain extending along the toe of the failure to a position further to the west. Unlike the CRA, there are no other scarps above the headscarp. A catastrophic failure at this location would almost certainly dam the Cascade River to a significant depth.

Franz Josef

Of particular note is the hillside above Franz Josef township, ~150 km northeast of the Cascade valley, which displays several of the structural characteristics identified at the Cascade and RT rock avalanches (Figure 4.10). At 700 m elevation there is a linear trough-like break in slope with a ~100 m-high scarp-like surface above and a bulging hillslope below. The Alpine Fault passes through the township and dips moderately beneath the hillside [e.g., Berryman et al., 1992; Barth et al., 2012]. The lateral distance to the town from the linear trough is 1300 m or 400 m measured from the base of the hill to the town. The mass of rock that could potentially fail at Franz Josef has comparable dimensions to the source of the Mt Wilberg rock avalanche [Chevalier et al., 2009]. The Mt Wilberg rock avalanche has an $L_{\text{max}}$ of ~2500 m (or a runout of about 1800 m measured from the base of the hillside) and an estimated deposit area of ~2 km² [Chevalier et al., 2009]; a comparable event at Franz Josef would completely devastate the town. There are several other localities along the Southern Alps rangefront where farmhouses and other accommodation are situated within kilometers of steep rangefront slopes.

4.3 Discussion

4.3.1 Coseismic triggering of the Cascade and Round Top rock avalanches

Several very large (> $10^6$ m³) deep-seated rock avalanches along the western Southern Alps and Fiordland rangefront (e.g., RT, Mt Wilberg, HBRR, John O’Groats) have been suggested to have been triggered by major Alpine Fault earthquakes [Wright, 1998b; Komp, 2004; Chevalier et al., 2009; Hancox and Perrin, 2009; Wood et al., 2011]. However, there are other sources of earthquakes in the region that should be given consideration. Historically, there have been about 20 M 5.0–6.2 earthquakes within a 50 km radius of the CRA, occurring primarily to the south, west, or along the Alpine Fault [GeoNet, 2012]. The largest historic
earthquake in this region thus far is a 1947 Ml 6.2 oblique-reverse event located 20 km southeast of the Alpine Fault (30 km south of the CRA) [Doser et al., 1999]. More distally, there have been several significant historic earthquakes associated with the Puysegur subduction zone offshore of Fiordland including the 1993 M 6.8 Secretary Island, 2003 M 7.2 Fiordland, and 2009 M 7.8 Dusky Sound earthquakes [GeoNet, 2012]. While the most significant landslides associated with the 2003 M 7.2 Fiordland were within less than 20–30 km of the fault rupture [Hancock et al., 2003], landscape disturbance evidence for a probable subduction-related event (or potentially a offshore Alpine Fault event) in AD 1826 extends as
far north as the central Alpine Fault [Norris et al., 2001; Wells and Goff, 2007; Howarth et al., 2012], suggesting the Puysegur subduction zone may be capable of larger earthquakes than have occurred historically and can generate widespread Modified Mercalli Intensity (MMI) up to X or even higher.

Cox et al. [2012a] mapped potentially active faults within the Southern Alps from around Haast (50 km northeast of the CRA) to Round Top and used fault geometry to model potential earthquake magnitudes. Although Cox et al. [2012a] find numerous potential sources of $M_w$ 5–7 events, and several faults capable of up to $M_w$ 7.1–7.5, according to their models the Alpine Fault remains the only source capable of $M_w$ 7.6 or greater earthquakes in this part of the Southern Alps. There are numerous active faults southeast of the Alpine Fault from Jackson Bay to Milford Sound (Chapter 2). Although the style of faulting is somewhat different in the general vicinity here as compared to further north, the fault lengths are comparable (3–13 km) and it seems a reasonable assumption that faults in the Southern Alps south of Haast (e.g., Chapter 2; Glade-Darrans Fault, Pembroke Fault and structures offshore Big Bay related to the Puysegur subduction zone) are capable of similar magnitude earthquakes. Despite other earthquake sources, Alpine Fault earthquakes have been demonstrated to be one of the most important drivers of erosion on the Southern Alps rangefront [e.g., Howarth et al., 2012]. A compilation of all historical earthquake-triggered landslides in New Zealand found that landslides with volumes $> 0.1$ km$^3$ occurred only in earthquakes of $M > 7.5$ and MMI IX or higher [Hancox et al., 2002]. The 2009 $M$ 7.8 Dusky Sound earthquake triggered widespread shallow landslides in an area of $\sim 2000$ km$^2$, but no deep-seated bedrock failures comparable to the $> 50$ very large to giant ($0.001$ to $\geq 0.1$ km$^3$) postglacial landslides (i.e., younger than c. 18 ka) known in Fiordland [Hancox et al., 2009]. The present study echoes the suggestion of Hancox et al. [2009] that the Alpine Fault and Puysegur subduction zone are likely the only two earthquake sources capable of triggering very large and widespread landslides in Fiordland and the southernmost Southern Alps.

Utilizing a radiocarbon age of 540 (+104 / -118) cal AD for a log entrained into the CRA deposit and a best guess age of the heartwood sample at the time of log entrainment of 120 yrs, yields a best estimate for the CRA of c. 660 AD. Though there is large uncertainty in this age, it overlaps well with the 642–727 AD Hk2 earthquake event recorded at Hokuri Creek 50 km to the southwest [U. Cochran pers comm., 2011; Berryman et al., 2012a] and a c. 630 AD megaturbidite event recorded in Lake Mapourika sediments 160 km to the northeast [Howarth, 2012]. The 642–727 AD event and the previous event at 110–231 AD (Hk3)
represent one of the longest recurrence intervals recorded in the continuous c. 8000 yr earthquake record at Hokuri Creek [Berryman et al., 2012a]. If the CRA failed during the c. 630 / 642–727 AD event, the long recurrence interval, > 200 km surface rupture, and failure of the CRA could be potential evidence of a particularly large (or great) Alpine Fault earthquake. Comparing rock avalanche ages to paleoseismology catalogs is one of the more definitive ways to determine if a prehistoric rock avalanche was triggered coseismically. Though there is large uncertainty in the c. 660 (+104 / -118) AD age of the CRA, it overlaps well with the 642–727 AD Hk2 earthquake event recorded at Hokuri Creek 50 km to the southwest [U. Cochran pers comm. 2011; Berryman et al., 2012] and a c. 630 AD megaturbidite event recorded in Lake Manapourika sediments 160 km to the northeast [Howarth, 2012]. The 642–727 AD event and the previous event at 110–231 AD (Hk3) represent one of the longest recurrence intervals recorded in the continuous c. 8000 yr earthquake record at Hokuri Creek (Berryman et al., 2012). If the CRA failed during the c. 630 / 642–727 AD event, the long recurrence interval, > 200 km surface rupture, and failure of the CRA could be potential evidence of a particularly large (or great) Alpine Fault earthquake. The present study obtains a 860 (+119 / -95) AD age for the RT rock avalanche (cf. 930 ± 50 AD age of Wright, 1998) by recalibrating the Wk-4914 radiocarbon date with the CALIB 6.0 program [Stuiver and Reimer, 1993] and the Southern Hemisphere Atmosphere correction curve [Reimer et al., 2009], and adding 95 ± 5 yrs (the number of growth rings into the log to the radiocarbon sample) to determine the date the sampled log was entrapped. This age overlaps with the Hk1 earthquake event (714–934 AD) identified at Hokuri Creek by Berryman et al. [2012], a megaturbidite event (965–887 AD) at Lake Paringa 85 km to the northeast identified by Howarth et al. [2012], and a previously identified, but poorly dated, event on the central Alpine Fault [Norris et al., 2001]. If the scarp-forming earthquake event at Hokuri Creek is the same event that triggered the RT rock avalanche, it may indicate the surface rupture of this earthquake extended more than 300 km along strike. Currently the only other major Alpine Fault earthquake with evidence for a surface rupture of this length is the most recent 1717 AD event [e.g., Wells et al., 1999]. Based on these correlations, it is proposed that the Cascade and RT rock avalanches were most likely triggered by major Alpine Fault earthquakes.

4.3.2 Large rock avalanches in the Southern Alps: an apparent paucity

The apparent paucity of very large rock avalanches along the Alpine Fault may be partly due to transient preservation of the deposits in a highly active landscape [e.g., Komp, 2005a,
characterized by ~12 m of rainfall annually [Tait et al., 2006] and calculated average erosion rates on the order of 10 mm/yr (Figure 1.3) [e.g., Cox et al., 2012a]. A catalog of 42 rock avalanches > 0.001 km$^3$ in a 10,000 km$^2$ area of the central Southern Alps found the observed frequency of one event per c. 250 yrs over the last c. 10,000 yrs was strongly biased by erosion [Whitehouse and Griffith, 1983]. All large rock avalanche deposits thus far recognized along or very near the Alpine Fault (e.g., Cascade, HBRR, RT, Mt Wilberg, John O’Groats, etc.) are not only post-last glaciation, but all have likely occurred in the last few thousand years. All of these deposits have been emplaced into active river valleys or floodplains and have been eroded to some extent by fluvial processes. The 1999 Mt Adams landslide (4 km southeast of the Alpine Fault) lost ~70% of its volume to the Poerua River by 2005 [Hancox et al., 2005]. The Late Holocene Mt Wilberg rock avalanche (at the Alpine Fault at the Wanganui River) has lost greater than 75% of its volume due to fluvial erosion [Chevalier et al., 2009]. Considering these case studies, the CRA deposit is remarkably well-preserved, presumably owing in part to its deposition in a wide, low gradient stretch of valley where the Cascade River has cut a confined river gorge through the deposit that localizes fluvial erosion. Further up the valleys, rock avalanche detritus deposited onto active glaciers [e.g., Cox and Allen, 2009; Allen et al., 2010] can be transported great distances and emplaced as lateral or terminal moraines [e.g., Reznichenko et al., 2012]; evidence for past rock avalanches may thus be obscured or erased. For example, little evidence remains of the 1991 Mt Cook rock avalanche [S. Cox pers comm., 2012], which was deposited onto the Hochstetter Glacier.

Furthermore, rock avalanches are commonly misinterpreted as glacial deposits in mountainous regions [e.g., Hewitt, 1999, and references therein]. Examples of previously misinterpreted large rock avalanche or landslide deposits in the Southern Alps/Fiordland include Green Lake [Hancox et al., 2009], The Hillocks [McColl and Davies, 2011], RT [Wright, 1998b] and Macaulay River [Burrows, 1972]; the CRA is yet another example. The Cascade valley is fortunate to have high lithologic diversity, which can help determine the nature and source of a deposit. The Cascade and HBRR rock avalanche deposits, for example, do not contain Haast Schist, which rules them out as glacial deposits in this catchment. Elsewhere in the Southern Alps, lithologies within catchments are less variable and this relationship is less diagnostic.

The HBRR rock avalanche highlights the issue that rock avalanches and glaciers can both produce remarkably similar deposit morphologies, especially when the surfaces have been eroded to some extent and are completely vegetated (Figure 4.1B). Through stereographic
aerial photographic interpretation and limited fieldwork, a contact has been drawn between what is believed to be the HBRR rock avalanche deposit and a glacial deposit to the southwest (Figure 1B). However, this distinction is subtle and based largely on a slight difference in surface morphology and vegetation as no deposit outcrops were found away from the river. Where deposits are thickly vegetated and outcrop is not present, the application of new technologies, such as airborne LiDAR or aeromagnetic data, may prove to be instrumental in delineating similar glacial and rock avalanche deposits in the future. The short survival time of rock avalanche deposits in the Southern Alps (often decades to centuries) and common misinterpretation as glacial features, indicates that their frequency and the hazard they pose may currently be significantly underappreciated.

4.3.3 Identifying future catastrophic failures

RSFs and sackungen are common throughout the Southern Alps and are often spatially associated with large (> 2.0 km\(^2\)) landslides; considered on a case-by-case basis, non-catastrophic RSFs and sackungen may be considered precursory phenomena to accelerated or catastrophic failure [e.g., Korup, 2005b]. Regardless of their mode of formation (e.g., post-glacial collapse or earthquake shaking), all sackungen are associated with existing failure surfaces that are capable of reactivation if strong ground accelerations can exceed the shear resistance imposed on these surfaces. These deep-seated surfaces exist due to gravitational instability, which means they are typically well-oriented to be reactivated catastrophically or otherwise. Key factors which may help identify sackungen and associated slopes that are good candidates for future catastrophic failure are (1) the length of sackungen and degree of segmentation (as greater length and less segmentation imply a more continuous surface at depth), (2) the amount of throw on a sackung (as more throw may indicate a better developed and frictionally weaker failure surface), (3) presence of tectonically-damaged bedrock at the base of the slope, (4) presence of bulging or oversteepened slopes below sackungen, and (5) degree of asymmetry of sackungen distributions. Asymmetric ridges (i.e., those where the elevation of the valley on one side of the ridge is lower than the other side and thus have asymmetric gravitational potentials) and associated lower-valley-facing sackungen may be particularly prone to deep-seated catastrophic failure as there are no opposing sackungen to interfere or compete with, as in symmetrical sackungen distributions (Figure 4.7). While the hillslope above Franz Josef (Figure 4.10) meets most of these conditions and forms one of the most compelling sites for further study in the area, there are
several other localities along the Southern Alps rangefront where farmhouses and other accommodation are situated within kilometers of steep rangefront slopes.

Large deep-seated rock avalanches pose a significant hazard through their large affected area, high velocity, long runouts, ability to create tsunamis, dam rivers, flood valleys, and result in catastrophic failures of rock avalanche dams [e.g., Robinson and Davies, 2013]. Despite an apparently low recurrence interval of voluminous ~2–5 km-long runout rock avalanches at the steep western rangefront of the Southern Alps, they remain a significant hazard. As the present study illustrates, catastrophic rock avalanches can initiate as relatively benign deep-seated bedrock failures. Identifiable precursory signs of failure should be fully considered in risk assessment catalogs and explored further.

4.4 Conclusion

Alpine Fault-related slope failures, including two very large catastrophic rock avalanches, are a prominent feature of the Cascade valley where it follows the fault for ~11 km. The 0.75 km$^3$ Cascade rock avalanche (CRA) is a well-preserved, catastrophic single-event failure with a source straddling the Alpine Fault. The CRA is interpreted to have failed coseismically with a large to great earthquake on the Alpine Fault based on (1) the clear structural relationship of the rock avalanche to the Alpine Fault and fault-damaged bedrock in its source area, (2) there are no known failures of comparable volume in the Southern Alps and Fiordland that did not have a coseismic trigger, (3) evidence that the Alpine Fault is likely the source of largest ground shaking and seismically-generated landscape driver in the region, and (4) the suggestion that very large (> 10$^6$ m$^3$) deep-seated catastrophic failures in Fiordland are associated with Alpine Fault earthquakes despite being less fault-proximal than the CRA. Utilizing a radiocarbon age of 540 (+104 / -118) cal AD and a 120 yr best guess for the radiocarbon sample at the time of entrapment into the CRA, this study obtains a best estimate of c. 660 AD for the age of the CRA and suggests a correlation to an Alpine Fault earthquake c. 630–727 AD.

The CRA deposit dammed the Cascade River for a distance of up to ~5 km upriver until the river cut a gorge through the deposit. Owing to the presence of the CRA deposit, a comparatively small volume failure (~10$^6$ m$^3$) is now sufficient to dam the Cascade River at the river gorge through the CRA deposit. The CRA mass was formerly a portion of a larger ~3 km$^3$ rock slope failure (RSF) associated with sackungen before the CRA catastrophically failed along a pre-existing failure surface (sackung). By analogy, two other rock masses in the
Cascade valley are identified as being primed for potential catastrophic failures of up to 2 km³.

This study demonstrates that although commonly benign, pre-existing slope collapse features like sackungen and RSFs should be considered potential precursors to catastrophic failure in risk assessments, especially when their distributions are asymmetric with respect to the ridge. The role of parent lithology and inherited fabric are subordinate factors in the generation of large slope failures on the rangefront of the Southern Alps. The common misinterpretation of large rock avalanches as glacial deposits and the short residence time of deposits indicates that their abundance, and thus risk, is currently underappreciated. They can have long runouts (~5 km) and large affected areas per event (~10 km²), which suggests that a significant component of the western Southern Alps rangefront, including populated areas like Franz Josef, may be at risk. Large catastrophic slope failures may occur relatively infrequently along the western rangefront of the Southern Alps, but the coseismic origin of many of them means that they will substantially increase the impact of large Alpine Fault earthquakes.
Chapter 5  Greater than 300 kyr of Landscape Evolution Preserved by Dextral-Normal Offset on the Southern Alpine Fault, New Zealand

Figure 5.0  Oblique aerial photograph looking west of southwest towards the Sara Hills (top center), Martins Bay (center distance) and Big Bay (far right) from a position above the Pyke River (not seen). The Skippers Range is at left and Lake McKerrow in the Hollyford valley is visible at upper left. The Alpine Fault (notice prominent west-side-up scarp at immediate bottom right corner) traverses diagonally towards the top left of the photograph. The glacially-sculpted hillslope at the bottom of the image has been dextrally offset ψ1450 m, at which time the glacier flowed from the Pyke valley into Big Bay (opposite to the Pyke River’s present course). Young marine sediments (Sara Formation) uplifted to 400 m elevation in McKenzie Creek (in saddle at center of photograph) were discovered by the present study. A paleo-reconstruction indicates that at their time of deposition the Sara Hills would have been an island, the Pyke and Hollyford valleys would have been fjords, and much of the southern Alpine Fault would have been beneath the Tasman Sea. The prominent hillslope break surrounding the Sara Hills may be glacial in origin (as previously proposed) or a paleo-shoreline. Photo taken by L. Homer (A7742A). Photo courtesy of GNS Science.

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Abstract

The landscape west of the Alpine Fault in South Westland is unique in that it is the only region adjacent to the fault in which bedrock-bounded river valleys are offset in a dextral-normal sense with a net uplift of the Australian (AUS) plate. This situation allows remarkable preservation of uplifted Quaternary marine sediments to elevations of ~600 m and a record of glacial deposits protected from subsequent glaciations by offset from their source catchment. I use paleontologic and stratigraphic age constraints on regressive marine sediments to determine fault-proximal AUS plate uplift rates of 2.2 (-0.4/+1.2) mm/yr and 2.5 (-0.4/+0.7) mm/yr, and overlying offset glacial deposits to constrain an Alpine Fault dextral slip rate of 29.6 (-2.1/+2.3) mm/yr, both over the last c. 300 kyr. In conjunction with other new and previous work, the results suggest the AUS plate uplift rates here have been constant over the last > 300 kyr and strike-slip rates on the southern Alpine Fault may have been relatively constant over the last > 3.5 Myr. These relatively constant strike-slip rates can then be used to estimate the timing of events such as glaciations, drainage reversals and catchment piracies. Along 100 km of the Alpine Fault, geomorphic features can be matched across an 8 km dextral offset, the largest known of its kind in the world. Fault offset reconstructions show that over the last > 500 kyr, the landscape best fits to a glacial event at 8 km of offset restored, or c. 270 ka (correlated to the Waimaunga Glaciation, Marine Isotope Stage 8). I use this relationship to suggest this glaciation was a major formative process in the establishment and sculpting of the modern river valleys here and presumably elsewhere in the Southern Alps.

5.1 Introduction

The Alpine Fault is one of the longest, straightest, least segmented and fastest slipping, active continental strike-slip faults in the world [e.g., Sutherland et al., 2007a; Molnar and Dayem, 2010]. Everywhere south of its junction with the Hope Fault (the southernmost of the through-going faults of the Marlborough Fault System) near the Toaroha River, the Alpine Fault is known to accommodate > 20 mm/yr of dextral slip (or 60–90% of plate boundary motion), whereas to the north it accommodates northward decreasing dextral slip rates of c. 12–3 mm/yr [Knuepfer, 1988; Norris and Cooper, 2001; Zachariasen et al., 2006; Barnes, 2009] (Figure 5.1A–B). This northernmost portion of the Alpine Fault presently accommodates very little of the plate boundary motion. Instead, slip is accommodated on younger (< 7 Ma) strike-slip faults of the Marlborough Fault System [e.g., Berryman et al., 1992; Little and Jones, 1998], including the Hope Fault, which presently accommodates c. 20 mm/yr of strike-slip and
forms the main linking structure between the Alpine Fault and the Hikurangi subduction zone [Van Dissen and Yeats, 1991; Langridge et al., 2003; Wallace et al., 2007; 2012]. The present study concerns itself primarily with rates of motion of the high-slip portion of the Alpine Fault, south of the Hope Fault.

It is well-established that dip-slip rates accommodated on the Alpine Fault vary along strike. Rates peak at greater than 12 mm/yr at Gaunt Creek (15 km northeast of Franz Josef), and reduce to zero near the Martyr River [Berryman et al., 1992; Norris and Cooper, 2001]. South of the Martyr River, the fault is dextral-normal with ubiquitous northwest-side-up fault scarps on steeply southeast-dipping fault planes observed both onshore and offshore indicating c. ≤ 3 mm/yr dip-slip rates [Berryman et al., 1992; Barnes et al., 2005; Chapter 2].

Strike-slip rates are notoriously more difficult to constrain on the Alpine Fault, in part due to the nature of preservation of horizontally offset markers in the erosive high rainfall rangefront environment dominated by dense bush and active river floodplains [e.g., Barth et al., 2012; Chapter 3]. Errors of ±5–6 mm/yr are not uncommon for these strike-slip rate determinations; such error bars encompass a substantial 30% of the c. 35 mm/yr Alpine Fault-parallel component of Australian-Pacific (AUS-PAC) plate motion [e.g., De Mets et al., 1994; Norris and Cooper, 2001]. Despite considerable overlap of most Alpine Fault strike-slip rate determinations (considering the errors) south of the Hope Fault, rates proposed by Sutherland et al. [2006] for the onshore Alpine Fault south of the Martyr River of 23 ± 2 mm/yr barely overlap with the 28 ± 5 mm/yr rate that is the average of three central Alpine Fault strike-slip rates reported by Norris and Cooper’s [2001], and rates of 27.2 (-3.0/+1.8) mm/yr and 31.4 (-3.5/+2.1) mm/yr rates determined from detailed bathymetry offshore Fiordland by Barnes [2009] (Figure 5.1B).

Together, these rates suggest that the southern and central Alpine Fault either (1) accommodates a relatively smooth along-strike southward increase in strike-slip rate, or (2) has a genuine along-strike zone of lower slip on the southern onshore Alpine Fault. The latter case would have particularly important implications for the long-term behavior of characteristic surface-rupturing earthquakes on the southern Alpine Fault, as it suggests this was a region through which slip failed to propagate so displacement therefore must be accommodated on other structures. This study contributes a well-constrained c. 270 kyr dextral slip rate for the southern onshore Alpine Fault and reexamines previously published data to resolve this disparity in along-strike strike-slip rates. Additionally, this chapter introduces new Alpine Fault-proximal AUS plate uplift rates, and provides constraints on
Alpine Fault-induced landscape evolution and the strength of past glaciations in South Westland.

5.1.1 Westland Coastal Plains & Hills province

For nearly 400 km along the Alpine Fault from near the Ahaura River in the north to where the Alpine Fault goes offshore near Milford Sound in the south, the land to the west of the Alpine Fault has a distinct geomorphology unique to New Zealand (Figure 5.1A–B; Figure 1.5; see Supplement 5 for a more detailed discussion). The four main features of this landscape are (1) well-developed alluvial fans which extend from where major rivers exit the Southern Alps to the Tasman Sea, (2) progradational coastal dune ridge complexes common from Hokitika to the Hollyford but particularly well-developed near Haast, (3) extensive Pleistocene glacial moraines and outwash gravel deposits to 300–400 m elevation, which bound alluvial fans and disrupt drainages to create an abundance of lakes and lowland swamps, and (4) bedrock “islands” of isolated hills or discontinuous ranges up to 1350 m in elevation, but typically ~600–900 m (Figure 5.1B; Supplement 5). The Westland Coastal
Plains & Hills province is unique in New Zealand in that it has a low density of faults and, with the exception of one small fault at Smithy Creek 5 km west of Franz Josef [Lund-Snee, 2011] and the Bald Hill Thrust south of Hokitika [Rattenbury, 1986], has no known active faults onshore (e.g., Figure 1.4). There appears to be very little Quaternary deformation of the onshore AUS plate south of the Ahaura River (see Supplement 5).

5.1.2 The onshore AUS plate in South Westland

South of the Arawhata River is the only place in the Westland Coastal Plains & Hills province where well-defined, bedrock-bounded river valleys that extend to the Tasman Sea are found west of the Alpine Fault (Figure 5.1C). These rivers are confined laterally in deeply incised valleys; combined with dextral-normal offset of the AUS plate this has led to remarkable preservation of Quaternary marine sediments and sub-aerial glacial deposits that presumably were widespread to the north along the Alpine Fault rangefront, but have been eroded or buried by subsequent fluvial and glacial erosion and sedimentation there. Because of the high lithologic diversity east of the Alpine Fault in South Westland (e.g., Figure 2.1B), AUS plate sediments can often be reasonably uniquely linked to sediment sources in catchments with headwaters on the PAC plate. Because onshore AUS plate deformation is highly-localized to the Alpine Fault and the fault forms a relatively linear and uncomplicated surface trace here, precise fault offsets can therefore be determined. A spectacular example is the middle Pliocene Halfway Formation exposed near the mouths of the Cascade and Hope rivers, which contains Fiordland-derived clasts indicating it has been dextrally offset > 100 km northeast of its source, requiring a strike-slip rate > 27 ± 4 mm/yr (possibly > 35 ± 5 mm/yr) over the last c. 3.5 Ma (assuming there has been no significant longshore transport of sediment between catchments) [Sutherland, 1994]. In contrast to the central Alpine Fault, tectonic geomorphology on the southern Alpine Fault suggests the AUS plate, which is relatively downthrown further north, has experienced Late Quaternary net uplift which continues to the present day (see Chapter 2).

In addition to providing very useful constraints on Alpine Fault strike-slip and uplift rates, this region can be used to estimate the timing of events such as glaciations, drainage reversals, catchment piracies, gorge incision, as well as the position of absolute sea level on this landscape through time (see also Supplement 5). The relative influence of these different processes on the geomorphology can be assessed. Since AUS plate lithologies and PAC plate drainage morphologies are comparable to further north, an understanding of the processes
responsible for landscape evolution and evidence for glaciations in South Westland can be effectively extrapolated north where recent Quaternary fluvial and glacial sedimentation has obscured or erased evidence of previous landscape and glaciations.

5.2 Glacial Deposits

5.2.1 New Zealand

As noted by Suggate [2004], the extent to which New Zealand glacial records have been preserved in the last c. 500 kyrs varies widely depending on the tectonic situation and the ease with which younger ice advances and fluvial erosion can obliterate evidence of older ones. Drill cores from offshore the east coast of the South Island record nine glaciations in the last 700 ka, but only evidence for the last four glaciations (back to c. 350 ka) is preserved onshore due to uplift and erosion [Suggate, 1990] (see also Figure S5.2 in Supplement 5 for MIS stage, name and age correlations of New Zealand glaciations and interglaciations). In North Westland, the area between the Hokitika and Grey Rivers is considered the type area for Middle to Late Pleistocene glaciations in New Zealand due to well-preserved morphostratigraphic relations between glacial outwash surfaces and uplifted beach terraces which allow correlations of glaciations back to marine isotope stage (MIS) 8 at c. 250 ka [Suggate, 1965, 1990, 2004; Suggate and Waight, 1999]. Many moraines and outwash surfaces in Westland terminate in cliffs at the coast, indicating their seaward extent was formerly greater.

5.2.2 South Westland

Glacial deposits are widespread on the AUS plate in South Westland, especially as extensive lateral moraines on the north side of major valleys and as eroded remnants of previously more extensive moraine and outwash surfaces up to elevations of 900 m (Figure 5.1C). Many of the higher elevation features are poorly-preserved, indicating their greater relative age. In general, glacial deposits south of Jackson Bay have received very little attention compared to further north. Regional reconnaissance work on glacial deposits was carried out by Turner [1930b] and Sutherland [1995]. Detailed work in this region has focused on the Cascade Plateau lateral moraine sequence north of the mouth of the Cascade River where $^{10}$Be cosmogenic nuclide ages between 117–14 ka have been recorded [Sutherland et al., 1995, 2007b], and at Hokuri Creek where the ages of Alpine Fault offset moraines were correlated against the North Westland sequence and used to determine strike-slip rates [Sutherland and Norris, 1995; Sutherland et al., 2006].
Although direct dating of glacial deposits in South Westland has been limited to the youngest and best preserved lateral moraines of the Cascade Plateau [Sutherland et al., 2007b], it is clear from the elevation of features, their degree of preservation, cross-cutting relationships, and the extent to which they have been offset by the Alpine Fault, along with correlation of moraine clast lithologies mentioned previously, that a partial record of several glaciations exists here (Figure 5.2, Supplement 5). Because dextral offset and net uplift of valleys on the AUS plate from their sediment sources on the PAC plate protect them from subsequent erosion by rivers and glaciers, this area offers a unique opportunity to study the way they record glaciations not as well-preserved elsewhere along the western rangefront of the Southern Alps. Using age constraints available, these offset glacial deposits can then be used constrain dextral slip rates of the Alpine Fault.

5.3 Regional 8 km Alpine Fault Dextral Offset

The largest (> 100 km²) area of outwash and moraine surfaces in South Westland is the Gorge River area which at present is not associated with any major valley which could have been the source of such a large glacier (Figure 5.1C; Figure 5.2C). Cross-cutting relationships, elevations, and relative preservation of features provide a record of multiple glacial events and indicate the oldest event involved a large distributary-lobed piedmont glacier with an extent greater than that of former glaciers in the Lower Cascade valley (Figure 5.3). Detailed

Figure 5.2  Oblique Aerial Photos of Glacial Features. (A) View of well-preserved lateral moraines of the Cascade Plateau looking west towards the Cascade River mouth (upper left). Note the decreasing preservation (or “freshness”) of lateral moraine crests towards the right of the picture as a function of increasing age. A high percentage of ultramafic clasts within this moraine, possibly supplemented by prehistoric fires, accounts for the stunted scrub present instead of thick rainforest. Even subtle features can be readily interpreted from stereoscopic aerial photos. Beneath this moraine is the Quaternary-aged Teer Formation (marine), which has been uplifted. The most laterally continuous moraines can be traced over 14 km from the coast to the highest point in the lower left foreground. Look direction is west-northwest. (B) View looking northeast towards the Cascade valley in the distance of the well-preserved paired lateral moraines of Low Creek. The lateral moraines are ~1 km apart. The Alpine Fault crosses diagonally from bottom right to top left. Note the lack of a source for the glacier that formed these moraines; they have been dextrally offset ~2900 m from the Jerry River headwaters effectively protecting them from being overridden by subsequent glaciations. Note also the subhorizontal lineaments on either side of the Low Creek lateral moraines; these are older moraine crests and glacial trim lines which extend out to the vast plateaux surrounding Gorge River and date to a time of ~8000 m of Alpine Fault offset. Look direction is northeast. (C) View looking west over the glacial plateaux of the Gorge River from a position near the center of the photo in B. Gorge River mouth at upper left. This vast region dates to a major glaciation that occurred when the Gorge River area was aligned with the main Pyke valley ~8000 m of Alpine Fault offset ago. Features are poorly-preserved but several cross-cutting relationships and surface elevations can be discerned with detailed stereoscopic aerial photography. Notice also the relatively small amount of fluvial incision which overprints the glacial deposits. Light-colored airstrip at left center is ~500 m-long. Look direction is northwest.
mapping with stereo-pair aerial photographs reveals moraine crests and glacial trim lines that indicate the source of the glacier was from the south (Figure 5.3). It is worth noting the similarities in glacial extent and morphologies between the Lower Cascade and Gorge river regions (Figure 5.1C).

The present study cut a digital hillshade of the region between Jackson Bay and Milford Sound along the Alpine Fault and used this to examine the relationships between the AUS and PAC plate topography during progressive restoration of 50 km of Alpine Fault dextral offset. This can be done effectively here because the Alpine Fault forms a narrow, straight trace with relatively pure strike-slip motion, the AUS plate crust here acts as a semi-rigid block, and the Alpine Fault accommodates more plate boundary displacement than all other faults in the region (e.g., Figure 2.1B; Chapter 2). Correlations (and misfits) of major valley margins and mapped glacial deposits were observed at every increment of dextral offset restored. Each offset occurred over a specific time period dependant on the corresponding Alpine Fault strike-slip rate. Along 100 km of the southern Alpine Fault, the landscape records a ~8000 m offset of geomorphic features, which is the best fit between AUS and PAC plate landscapes at all times (0–50 km offset), including at the Last Glacial Maximum (LGM) and presently (Figure 5.4; cf. Figure 5.1C). To the author’s knowledge, this is the largest regional geomorphic offset recorded anywhere in the world.

Once this 8000 m offset is restored, all major AUS plate drainages except the Lower Cascade align with the next PAC plate valley to the south (e.g., McKenzie Creek/Big Bay with the Hollyford; Lower Hollyford with Upper Kaipo; John O’Groats/Wolf Tablelands with Milford Sound). The prominent right bend in the Cascade valley disappears and the widespread glacial surfaces in the Gorge River area become aligned with the main Pyke valley. Mt McLean moves to the southwest side of the Arawhata River outflow. Lateral moraines of the Wolf Tablelands (discussed in detail subsequently) restore to the north side of Milford Sound, consistent with their clast composition. Although the Gorge River deposits were not examined in detail in the field, beach boulders at the coastline between the Gorge and Spoon rivers are consistent with the moraine being sourced from the Pyke valley. Hokuri Creek is the most notable misfit, which may indicate it formed more recently than the other features. Similarly, the Dry Awarua area may have formed after the Gorge River area was offset from the main Pyke valley. Attempting to align the major AUS valleys one further to the south (i.e., ~16 km offset), yields a poor correlation.
Figure 5.3  Detailed interpretation of glacial features in the Gorge River area. Note the two-lobed nature of the highest and oldest surfaces comparable to the pattern observed for LGM deposits at the terminus of the Cascade River valley. The main outlet formerly was to the north in the Spoon River area. Moraine crests and trim lines indicate the red, orange, pink and purple surfaces all relate to glaciations sourced from the main Pyke valley (~8000 km offset), while the cross-cutting green surfaces date to a much younger glaciation (at ~2900 km offset). Uplift and abandonment of the plateaux defining a former glaciated valley floor has driven the Gorge River to cut a steep and narrow gorge through underlying bedrock at its river mouth.
This remarkable alignment of major glaciated valleys across the fault and widespread glacial deposits dating to this time (especially in the Gorge River area), provide compelling evidence that a synchronous glacial event was of major importance in the sculpting and establishment of the modern river valleys in South Westland during the Plio-Pleistocene. Although evidence for this glaciation is not as well-preserved elsewhere in Westland, I suggest this glaciation may have had a similar impact further north and elsewhere in New Zealand. Because of the difficulty in directly dating glacial deposits beyond the age limits of radiocarbon methods (c. 60 kyrs) in this environment, ages are estimated using paleontological and stratigraphic constraints from underlying marine deposits and by correlation with the timing of local and global paleo-climate records. In subsequent sections, I use these ages not only to correlate glaciations, but also provide well-constrained medium to long-term AUS plate uplift rates and Alpine Fault dextral slip rates.

5.4 Marine Deposits

Young fossiliferous Quaternary marine sediments are the most conspicuous evidence of AUS plate uplift in the region. These occur in four main locations at elevations from 50 m to 590 m and are everywhere overlain by glacial deposits, which seem to be a requirement for their preservation (Figure 5.1C). The Teer Formation, exposed in coastal outcrops beneath the Cascade Plateau at ~80 m elevation, has been described in detail by Sutherland et al. [1995] who interpreted the shallow marine fossiliferous silts as fjord sediment correlative with the oldest glacial moraine deposits at the Cascade Plateau (crudely estimated at 0.9 ± 0.4 Ma). A similar outcrop of fossiliferous silts exposed in the Spoon River at ~50 m elevation may be a correlative of the Teer Formation [Sutherland, 1995]. Although the present study relies on

Figure 5.4 8 km of dextral Alpine Fault offset restored. The best fit of the major valleys on opposing sides of the Alpine Fault over the last c. 500 kyr (including today) occurs at ~8000 m of dextral offset is restored. An age for this offset of about 270 ka can be deriving using the constraints placed by the Wolf Tablelands and Wolf Formation. Compare this figure to Figure 5.1C. At this time the Cascade is the only major AUS plate valley not offset from its modern source; all other major AUS plate valleys align with the next major drainage on the PAC plate to the south. In particular, note the position of the Wolf Tablelands as a lateral moraine north of Milford Sound and the Gorge River glacial surfaces aligning with the Pyke valley which hosted a large glacier flowing from the Upper Hollyford. The comparably small volume of glacial sediments associated with the Lower Hollyford valley suggests a drainage divide existed between the Lower Hollyford and Upper Hollyford at this time (i.e., the modern Upper Hollyford used to be the headwaters of the modern Pyke valley). Hokuri Creek and the Dry Awarua River area presumably post-date this time. Note that for simplicity modern sea level is shown but actual relative sea levels on the AUS plate would be ~500 m higher in places. Northwest illuminated bathymetry hillshade (colored) derived from 75 m DEM data courtesy of NIWA.
observations of the Teer Formation by Sutherland et al. [1995], *Talochlamys* (scallop) samples that Sutherland lodged in the New Zealand Fossil Record Electronic Database (FRED) were also analyzed for amino acid racemization for comparison with sites focused on in detail in the present study. Two new Quaternary marine formations are introduced in the present study. A summary of the paleontology and nature of the key sites is given in Table 5.1; more detailed paleontology is given in the Appendix.

### 5.4.1 Sara Formation

A new unit of Late Quaternary fossiliferous marine silts and sands, herein named the Sara Formation, was identified in the course of fieldwork during the present study at an elevation of ~400 m and only 100 m laterally from the Alpine Fault at McKenzie Creek southeast of Big Bay. Here, two actively eroding slip gullies provide excellent outcrop of a ~30 m section of laminated fossiliferous marine silt and sands overlain by a > 50 m section of morainic deposit. The top of the morainic deposit is contiguous with a widespread glacial surface extending across Jamestown Saddle towards Hokuri Creek to the southwest and it is suggested the underlying Sara Formation may extend similarly. Although fragile and commonly shattered in situ, macrofossils are well-preserved. These have fine ornamentation, color, patterns and preserved mother-of pearl (Figure 5.5A). Well-preserved woody debris is also common throughout the deposit; pieces up to 6 cm long have been positively identified as angiosperm wood, possibly a species of nothofagus (beech tree) [J. Bannister pers. comm., 2012]. Macrofossils are distributed throughout the unit, but are especially common within discrete horizons, potentially indicating shell drifts (Figure 5.5B).

Meter-thick horizons of sub-angular, locally-derived Webb Gneiss are distributed through intervals of 2–5 m and presumably indicate deposition by slips from a nearby steep fjord wall into the sea (Figure 5.5C). Subrounded-rounded glacial dropstones up to 0.5 m in diameter indicate the presence of floating ice; they are present throughout the unit, but increase in abundance upsection (Figure 5.5D). The contact of the Sara Formation with the overlying morainic deposits was only able to be observed from a distance, but it did not appear to comprise a major angular unconformity or weathered surface (which could indicate a period of non-deposition). Palynology indicates cool to cold temperatures and that forest trees were nearby at the time of deposition. The foraminiferal assemblage indicates deposition in sheltered normal marine salinity conditions, consistent with the macrofossil assemblage that
indicates deposition in less than 10 m of water with the presence of a hard rock substrate nearby.

5.4.2 Age and paleo-topographic reconstruction of the Sara Formation

The presence of the coccolithophore *Emiliana Huxleyi* in the Sara Formation samples from McKenzie Creek indicates a depositional age of < 290 ka [e.g., Gradstein et al., 2012]. Amino acid racemization analyses from the site give a D/L value of 0.400 ± 0.07, which, assuming a diagenetic temperature of 8.7 °C (best approximation for mean annual temperature) and correlation to well-dated sequences from the Wanganui Basin (North Island) and eastern Australia, yields a depositional age estimate of 191 ± 34 ka for the unit (see Supplement 5) [C. Murray-Wallace pers. comm., 2011]. Using this age estimate of 191 ± 34 ka, a eustatic sea level over this time between -20 m and -120 m [Siddall et al., 2006], a water depth at time of deposition of 0–10 m, and a modern elevation of 400 (-3/+5) m, I calculate a local AUS plate uplift rate of 2.2 (-0.4/+1.2) mm/yr.

Together the observations of upward increasing abundance of dropstones and palynological indications of a cool to cold climate suggest the Sara Formation was deposited as a regression towards sub-aerial glacial conditions, associated with steady uplift and a strengthening glaciation, which would have coevally driven sea level lower. For these delicate marine sediments to be preserved, it is suggested the overlying morainic deposits would have had to be deposited shortly following deposition of the marine unit. On the basis of the 191 ka estimate, the Sara Formation is assigned to early Marine Isotope Stage (MIS) 6, which indicates the overlying glaciation is related to the Waimea Glaciation (although the top of the morainic deposit may have been altered by more recent glaciations; see also Supplement 5).

The presence of the Sara Formation in the bottom of the valley between the Skippers Range and Sara Hills (indicating topography has not changed significantly since its deposition), in conjunction with age and water depth estimates, allows an accurate paleo-reconstruction of the landscape to be made (Figure 5.6). This reconstruction indicates the McKenzie Creek site was deposited in a shallow beach setting adjacent to a rocky shoreline at the mouth of the Lower Hollyford/Hokuri fjord. The site would have been partially sheltered from the open sea by the Sara Hills, which at the time were likely an island.
Figure 5.5    Quaternary marine sediments. A-D are photos of Sara Formation exposures at McKenzie Creek. (A) Remarkable preservation of shells (fine ornamentation, color and patterns preserved) and woody debris (below eraser) in uplifted marine sediments at 400 m elevation near McKenzie Creek. View looking north. (B) Example of laminated marine sediment with shell and carbonaceous-rich horizons. View looking northwest. (C) Meter-thick horizons of locally-derived Webb Gneiss likely indicate slip detritus from a nearby wall of what would have been a shallow fjord entrance. View looking northwest. (D) Well-rounded dropstone of Mistake Diorite sourced from either Hokuri Creek or the Lower Hollyford valley. Notice the soft sediment deformation of silt layers beneath the weight of the boulder providing unequivocal evidence for a dropstone origin. View looking southwest. (E) Asymmetric climbing ripple marks occur at the top of the marine Wolf Formation at Wolf River below morainic till. An upward coarsening sequence from silt to sand, in conjunction with upward increasing current influence and increasing dropstones, indicates a regression towards the overlying sub-aerial morainic till. Sample is out of situ. (F) View of the slip exposure at Madagascar Creek looking south. Blue-grey marine silts of the Wolf Formation (here 30 m thick) are visible at lower right overlain by 5 m of unfossiliferous brown coarse sand exposed above the dashed line and a 35 m section of morainic till of the Wolf Tablelands. Clast lithologies of the till are consistent with a source from Milford Sound.

5.4.3 Wolf Formation

Fossiliferous marine sediments were observed by Turnbull et al. [2010] as float in several slips below the Wolf Tablelands morainic till, but until the present study had not been observed in outcrop. This unit is herein named the Wolf Formation. The Wolf Formation is only observed as a 15–30 m-thick section underlain by Greenland Group metasedimentary basement and overlain by morainic till of the Wolf Tablelands at elevations between 400 and 590 m. Although it is only observed in outcrop in western and northern-facing margins of the Wolf Tablelands (Figure 5.7), it is presumed to underlie the entire gently southwest-dipping plateau, including potentially the Table Land in the south and Mt McGulsh (684 m) in the north.

The best section visited is at the informally named Madagascar Creek, the creek immediately southwest of the Wolf River. Here, fine blue-grey marine silts have been deposited on the unconformable contact with Greenland Group basement; this surface appears to have had little topographic expression and no lag deposit of Greenland Group clasts is present. Marine macrofossils are present through the ~30 m section of marine silts and sands, but are especially abundant 9–12 m and 15–17 m above the base of the deposit, and especially rare in the upper ~5 m of the section. Fossils within the outer 5–10 cm of the outcrop surface have been decalcified by surface weathering, commonly only molds remain. However, deeper excavation reveals well-preserved fauna with fine ornamentation, mother-of-pearl, and occasionally color. Faint ~1 cm-wide parallel laminations are present throughout; these dip shallowly to the southwest consistent with the overlying morainic till surface. Subrounded-
### Table 5.1: Summary of Uplifted Marine Sediment Paleontology

ID numbers refer to the New Zealand Fossil Record Electronic Database (NZ FRED). Samples are lodged at the University of Otago. See Supplement 5 for detailed discussion of amino acid racemization (AAR); note D/L values are pooled by formation and not ID. See Appendix for detailed paleontological reports. Depth values are pooled by formation and not ID. See Table 5.1 for more details.

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<th>AAR D/L Value</th>
<th>Depth (m)</th>
<th>Macrofossils</th>
<th>Pollen</th>
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<td>0.398 (+0.03 / -0.02)</td>
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<td>r Miocene; m dom s.f.; 0 – 10</td>
<td>0 10 30</td>
<td>D39/f0095 * Wolf River 3 – 4</td>
<td>590</td>
<td>0.448 (+0.07 / -0.04)</td>
<td>0.431 (+0.06 / -0.05)</td>
<td>&lt; 290</td>
<td>r (1.8 – 0.55); m dom 10 – 50</td>
<td>10 50 200</td>
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<td>0.431 (+0.06 / -0.05)</td>
<td>&lt; 290</td>
<td>r (1.2 – 0.55); m dom 10 – 50</td>
<td>10 50 200</td>
<td>D39/f0097 * Wolf Madagascar Creek 15 – 17</td>
<td>560</td>
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<td>0.431 (+0.06 / -0.05)</td>
<td>&lt; 290</td>
<td>r (1.8 – 0.55); m dom 10 – 50</td>
<td>10 50 200</td>
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*Same stratigraphic unit so age analyses pooled*
Figure 5.6  Sara Formation landscape reconstruction (A) Present day hillshade derived from 15 m LINZ DEM data in the vicinity of McKenzie Creek showing Alpine Fault-induced drainage divides and major watercourses (blue) draining to the Tasman Sea. (B) Reconstruction at ~5200 m offset restored (c. 190 ka at 27 mm/yr dextral strike-slip). The expression of sea level on the landscape is estimated utilizing water depth estimates for the Sara Formation, uplift estimates at Lake McKerrow by Norris and Cooper [2001] and the presumption that AUS plate uplift rates decrease away from the Alpine Fault. A prominent seaward-dipping slope break that encircles much of the Sara Hills is either a glacial trim line or a paleo-shoreline. All the major valleys would have been fjords at this time and much of the Alpine Fault in the area would have been submerged. Dashed regions denoted as “Potentially land?” may have been eroded since this time. Yellow-infilled circles denote locations of drainage divides induced by dextral-normal motion on the Alpine Fault.
Figure 5.7  Wolf Tablelands. Extent of the ~600 m elevation Wolf Tablelands, a shallowly southwest-dipping plateau of morainic till with lateral moraine crests arcing towards the southwest, and underlying Late Quaternary marine Wolh Formation. Locations of the Wolf River and Madagascar Creek sites discussed in the text are denoted by stars. Poor preservation of features, moraine clast composition and underlying Wolf Formation indicate the Wolf Tablelands has been offset from a Waimaunga-aged glacier in Milford Sound.

rounded glacial dropstones to 1.5 m-across increase upsection and are composed of Jagged Gneiss or Anita Ultramafics (rare) consistent with floating ice from the Milford Glacier.

A ~5 m section of subangular medium-grained lithic sand separates the fossiliferous marine silts and sands below from a ~30 m section of morainic till (with minor glacial outwash) above (Figure 5.5F). While Jagged Gneiss and Milford Orthogneiss dominate the clast lithology, Anita Ultramafics mylonite clasts form a distinct ~5–10 % component of the moraine deposit. The Anita Ultramafic rocks are not abundant in outcrop north of Milford Sound, the moraine composition is most consistent with a source from the Milford glacier. The Wolf Tablelands has poorly-preserved arcuate lateral moraine crests, which can be readily mapped using stereographic-paired aerial photographs (Figure 5.7). The moraine crests have southward arcing traces and define a fanning distribution from east-west crests in the north to northeast-southwest crests in the south. This pattern is comparable to lateral moraine crests preserved on the north side of major glacial valleys (e.g., Cascade, Gorge River area between the Spoon and Hope rivers (Figure 5.1C). The degree of preservation of the surface
morphology of the *Wolf Tablelands* lateral moraine, height of the deposit, and the amount of
creek incision into its northwestern side, all suggest the deposit is one of the oldest preserved
in the area. Together these observations indicate the *Wolf Tablelands* is a remnant of a
formerly more extensive lateral moraine formed on the north side of a glacier from Milford
Sound. The regionally-preserved Alpine Fault strike-slip offset recorded by all of the major
river valleys (especially prominent at the Cascade and Pyke/Gorge), precisely constrains this
offset to be ~8000 m (Figure 5.4).

As the region of preserved Wolf Formation is presently a topographic high, paleotopography
at its time of deposition is more difficult to directly reconstruct than for the Sara Formation.
Macro- and microfauna are consistent with water depths of about 10–50 m for the base of
the Wolf Formation, while microscopic “beach-like” charcoal, a lack of nannofossils, and
climbing ripples are consistent with a shallow beach setting for the top of the deposit.
Palynology indicates a dominance of daisy, grass, shrub and herb pollen, suggesting there was
no extensive forest near the site. The macrofauna from the stratigraphically highest portion
of the Wolf Formation is epifaunal (lived attached to rocks or shells), which indicates
deposition into the site from a nearby rock face. Paired valves of *Tabulamys gemmulata* and
*Pratulum pulchellum* in the lower portions of the Wolf Formation indicate a calm depositional
environment. The foraminiferal assemblage indicates deposition in sheltered normal marine
salinity conditions; species are indicative of ~50–200 m water depths and several well-
preserved, but reworked extinct species indicate reworking from a ~1.8–0.8 Ma source. The
lack of wood and proximal slip detritus (as is present in the Sara Formation), indicates a less
sheltered depositional environment for the Wolf Formation. There is no marine unit exposed
on the edge of the lower elevation glacial surface at the head of the Kaipo Slips to the
southeast, which together with the truncated southeast edge of the *Wolf Tablelands* lateral
moraine crests, indicates this lower glacial surface is younger (Figure 5.7).

Like the Sara Formation, the Wolf Formation is a regressive sequence. The sequence
transitions from fine silt at the base to medium-grained lithic sand at its top. Climbing ripples
(Figure 5.5E) near the top of the sequence at Wolf River indicate it was deposited in a
shallow environment where the influence of current was important, and while sediment input
was increasing. Microscopic “beach-like” charcoal is found at the top of the *Madagascar Creek*
marine section, but not below. Also like the Sara Formation, the cool to cold climate
indicated by palynology, the upsection increase in dropstones, the fact this is a regressive
sequence, and the lack of a discrete unconformity between the marine sediment and overlying
35 m section of sub-aerial glacial deposits, suggest the Wolf Formation is an early glaciation
deposit overlain by mid-peak glacial deposits of the same glaciation. A more-or-less
continuous section between the Wolf Formation and overlying sub-aerial glacial deposits
indicate a rapid fall in sea level due to the strengthening glaciation, in conjunction with local
uplift of the AUS plate. The immediate burial of the fragile marine sediments by glacial
moraine has allowed the deposit to be preserved.

5.4.4 Age and tectonic implications of the Wolf Formation

The presence of the coccolithophore *Emiliana Huxleyi* indicates an age of < 290 ka [e.g.,
Gradstein et al., 2012] for the Wolf Formation, a very important age constraint for this deposit.
Realistic Alpine Fault strike-slip rates in the region (~21–33 mm/yr) [Norris and Cooper,
2001; Sutherland et al., 2006; Barnes, 2009] require the overlying glacial deposits (offset 8 km) to
belong to either Marine Isotope Stage (MIS) 10 (Nemona Glaciation, c. 340 ka) or MIS 8
(Waimaunga Glaciation, c. 270 ka) [Jouzel et al., 2007; Barrell, 2011]. Since the Wolf Formation
must be less than 290 ka, the overlying glacial deposits are attributed to the Waimaunga
Glaciation (MIS 8) and are assumed to be peak glacial in age.

5.4.4.1 Australian plate uplift rates

Because of these relationships, the age of the Wolf Formation can be tightly estimated to be
between 290–270 ka. Vertical uplift of the Wolf Formation is estimated using an present
elevation of 590 m with water depth constraints from the macrofauna and foraminiferal
assemblages to have been 10–200 m, and eustatic sea levels at the time of ~80 to -20 m
[Siddall et al., 2006] to obtain 700 (-80/+170) m. Thus, using a vertical uplift of 700 (-80/+170) m and an age of 280 (±10) ka, I calculate a c. 300 kyr Alpine Fault-related AUS
plate local uplift rate of 2.5 (-0.4/+0.7) mm/yr. This rate is comparable to a 2.2 ± 0.2 mm/yr
AUS plate uplift rate at Lake McKerrow (at a locality a similar distance from the Alpine Fault
as the Wolf Tablelands) derived from c. 15.6 ka uplifted intertidal shell beds [Norris and Cooper,
2001]. For comparison, the same c. 15.6 ka uplifted intertidal shell beds on the Pacific plate at
Lake McKerrow yield an uplift rate of 1.6 ± 0.3 mm/yr, consistent with geomorphic
observations of net uplift occurring on AUS plate. Considered together, these rates suggest
uplift of the AUS plate immediately adjacent to the southern Alpine Fault has occurred at
relatively constant rates over the last c. 300 kyr.

Apatite and zircon fission track data from a fault-perpendicular transect from the Alpine
Fault at the Arawhata River Bridge to Jackson Bay yield uplift rates of 0.7–0.8 mm/yr along
the coast and ~1.0–2.0 mm/yr adjacent to the fault if 5 Ma is taken as the time uplift started [Kamp et al., 1992]. Sutherland et al. [1995] determine similarly low (< 1 mm/yr) uplift rates at the Cascade River mouth based on estimated ages and water depths of the Teer and Halfway formations. Studies further north on the AUS plate adjacent to the central Alpine Fault have also found uplift rates of < 1 mm/yr along the coast and ~2.0 mm/yr adjacent to the fault [Bull and Cooper, 1986; Seward and Nathan, 1990; Cooper and Kastro, 2006]. This apparent spatially uniform pattern of AUS plate uplift rates along-strike of the central and southern Alpine Fault is remarkably consistent despite dramatic along-strike changes in PAC plate uplift rates [e.g., Norris and Cooper, 2001], consistent with the idea that the onshore AUS plate is deforming as a semi-rigid northwestward-tiling crustal block.

5.4.4.2 Strike-slip rates

Using a 8000 (-150/+520) m dextral offset constrained by restoration of the Cascade valley, and an age of 270 (-3.0/+15) ka for the Wolf Tablelands lateral moraine estimated from EPICA Dome C ice core 800 kyr deuterium data [Jouzel et al., 2007], I calculate a c. 300 kyr Alpine Fault dextral slip rate of 29.6 (-2.1/+2.3) mm/yr for the onshore southern Alpine Fault from Jackson Bay to Milford Sound. Sutherland [1994] and Sutherland et al. [1995] used the reasonably well-constrained age of the Halfway Formation conglomerate (3.6 ± 0.5 Ma) exposed beneath the Cascade Plateau and the clear Fiordland provenance of its clasts (requiring a minimum strike-slip offset of 95–100 km) to determine a minimum Alpine Fault dextral strike-slip rate of ≥ 27 ± 4 mm/yr over the last c. 3.5 Ma; this is consistent with my estimated rate over the same region. However, these two rates are higher than the 23 ± 2 mm/yr strike-slip rate estimated by Sutherland et al. [2006] over the same region based on offset hillslopes and moraine crests inferred to be 79–18 ka old. Taken at face value, these data suggest the dextral slip rate is decreasing through time over this portion of the Alpine Fault. Because of this implication, during the present study I have reexamined all of Sutherland et al. [2006]’s offset features and reconsidered their methods.

5.5 Re-examination of Southern Alpine Fault Dextral Slip Rates

5.5.1 Sutherland et al. [2006]

The present study calls into question the apparently well-constrained Alpine Fault dextral slip rates of 23 ± 2 mm/yr calculated by Sutherland et al. [2006] based on 12 offset hillslopes and correlation of offset moraines from the Martyr River in the north to the John O’Groats River

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in the south. As discussed in Chapter 2, these rates are lower than those recorded on the central Alpine Fault to the north and from offset of features visible in offshore bathymetric data [e.g., Norris and Cooper, 2001; Barnes, 2009]. A true deficit in strike-slip rate on the southern onshore Alpine Fault has important tectonic implications; it may indicate this is a region through which some major earthquakes fail to propagate, have lower net coseismic displacements (with some partitioning occurring onto other parallel faults), and/or accommodate a component of fault creep (see discussion in Chapter 2).

The offset features are glacial in origin (glaciated valley hillslopes and moraine crests). None of the features they use to make these estimates have been directly dated. Instead, correlations to New Zealand (specifically cosmogenically dated moraines on the Cascade Plateau), South Pacific and global records were used to give ages of 18, 22, 58, 79 ka. I do not question this approach but point out that there are conservative uncertainties in ages estimated by this method of about 5% of the age for LGM features. Offset glaciated hillslopes are considered for catchments of many different sizes and geometries and may potentially reflect different ages of glacial retreat. The following discussion considers the offset hillslopes. A new correlation of the offset moraines examined by Sutherland [1995], Sutherland and Norris [1995], and Sutherland et al. [2006] is discussed subsequently.

For all offsets, Sutherland et al. [2006] assumed there was no subsequent erosional modification of the features, which is a valid assumption where moraine crests can be correlated (four examples) but questionable for the edges of offset hillslopes (eight examples). In the absence of linear markers on the offset hillslopes, the authors chose control points on representative 20 m contour lines which were then horizontally extrapolated across the fault, employing a highly interpretive bend in the hillslope where there has been erosion at the fault (e.g., their SW Lake McKerrow offset). This method requires large ~300–700 m lateral extrapolations (between control points) across the fault on ~400–500 m offsets. Furthermore, their offset determinations assume pure strike-slip offset which is a gross simplification. Everywhere south of the Martyr River, Alpine Fault slickenlines plunge 7–10° to the southwest on steeply southeast-dipping fault planes consistent with up-to-the-northwest scarps observed (Chapter 2). A single event Alpine Fault displacement of 7.5 ± 0.5 m dextrally and 1 m up to the northwest at Hokuri Creek [Berryman et al., 2012a] is consistent with this. Thus for a 450 m horizontal offset, the associated vertical offset is expected to be 60–80 m, or three to four 20 m contours.
During the present study I have examined all of *Sutherland et al.* [2006]’s eight glaciated hillslope offsets by taking digital images of topographic maps, 15 m DEM-derived hillshades, and aerial orthophotos, cutting them along the Alpine Fault in a photo-editing software (Photoshop), and then manually restoring the offset features to their best fit (while considering a component of AUS-plate-up dip-slip which would result in misalignment of sloping surfaces). At the Martyr River *Sutherland et al.* [2006] correlate a feature they interpret as a moraine crest to the Cascade valley edge, which is in fact a portion of the more recent headscarp of the Cascade rock avalanche (Chapter 4). Their apparent hillslope offsets of 447 and 475 m at the Jerry River may actually be related to an older ~2900 m offset (see below). In summary, the present study finds issue with the offset measurement method, assumption of pure strike-slip offset, moraine age correlations and the offset features restored during the study by *Sutherland et al.* [2006]. Combined with on-the-ground examination of many of the hillslopes, I conclude that a single hillslope offset surface does not provide a reliable slip rate determination in this environment and that statistically pooling such offsets does not overcome systematic uncertainties. I propose that the ± 2 mm/yr errors on *Sutherland et al.* [2006]’s strike-slip rate of 23 ± 2 mm/yr are in fact likely to be much larger than their statistical estimation due to this uncertainty about key assumptions of feature age and offset. As single offset markers of surfaces across river valleys are not always reliable and can usually only be used to give minimum or maximum strike-slip rates [e.g., discussion in Norris and Cooper, 2001], I suggest I have provided a more reliable determination of Alpine Fault dextral slip rate over the same time and area in the current study.

### 5.5.2 Dextral slip rates derived from this study

As single offset markers of surfaces across river valleys are not always reliable and can usually be only used to give minimum or maximum strike-slip rates [see also Norris and Cooper, 2001], the present study instead focuses on a smaller subset of offsets with paired or multiple displacements (Table 5.2). Aerial photography, digital hillshades and topographic maps were all employed to constrain offsets. Ages of offset features were estimated from glacial correlation to EPICA Dome C ice core 800 kyr deuterium data [Jouzel et al., 2007] (Figure S5.2) or ages introduced by *Sutherland et al.* [2006].

Offset of the southwest valley hillslope of the Kaipo valley (utilized in *Sutherland et al.*, 2006) is reconsidered in conjunction with hillslope reconstruction near *Butchers Creek* on the north side of the Kaipo valley (Figure 5.8). Together these imply a best-fit offset of ~570 m. It is
Table 5.2

Summary of Key Glacial Features. "Offset" refers to the best-fit dextral (horizontal) offset, while "min" and "max" refer to constrained minimum and maximum values. Rates are estimated from geometric and stratigraphic constraints from the present study. See also Figure S5.2.

<table>
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<th>Feature</th>
<th>NZ Location</th>
<th>Glacial History</th>
<th>Age (ka)</th>
<th>Lat</th>
<th>Lon</th>
<th>Age (°)</th>
<th>Lat</th>
<th>Lon</th>
<th>Detrital Slip Rate (mm/yr)</th>
<th>Min</th>
<th>Max</th>
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</thead>
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<td>8 MIS</td>
<td>Late Otira Kaipo</td>
<td>18</td>
<td>2</td>
<td>2</td>
<td>2</td>
<td>3</td>
<td>11</td>
<td>15</td>
<td>29.6</td>
<td>31.7</td>
</tr>
<tr>
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<td>6? MIS</td>
<td>Early Otira Pyke</td>
<td>58</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>3</td>
<td>2</td>
<td>17.5</td>
<td>31.5</td>
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<tr>
<td>Early Otira Hokuri Moraine</td>
<td>58</td>
<td>Early Otira Pyke</td>
<td>58</td>
<td>5</td>
<td>5</td>
<td>5</td>
<td>3</td>
<td>2</td>
<td>22.7</td>
<td>33.1</td>
<td></td>
</tr>
<tr>
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<td>Late Otira</td>
<td>138</td>
<td>7</td>
<td>9</td>
<td>9</td>
<td>3</td>
<td>1</td>
<td>25.0</td>
<td>31.5</td>
<td></td>
</tr>
</tbody>
</table>

Age is estimated from geometric and stratigraphic constraints from the present study. See also Figure S5.2.

Table 5.2
Figure 5.8 Kaipo River offsets. LINZ topographic map of (A) present day and (B) ~570 m dextral offset restored (~LGM) showing two post-LGM hillslope offsets. This offset is considerably larger than the ~440 m offset estimated by Sutherland et al. [2006] from contours extrapolated across the southwest valley hillslope offset.
Figure 5.9    Pyke Corner glaciated hillslope. LINZ topographic map of (A) present day and (B) ~1450 m dextral offset restoration of a glaciated hillslope between McKenzie Creek and the Pyke River. The latter is the most recent time a glacier flowed from the Pyke River past this corner into Big Bay. The greater amount of creek incision on the western flanks of the corner potentially indicates ice has been free from this hillslope for a longer period of time than on the north to northeastern flanks.
Figure 5.10  Low Creek lateral moraines. Aerial imagery of (A) present day and (B) ~2900 m dextral offset restored of two sets of well-preserved paired lateral moraines in the Low Creek and Jerry River region. These moraines cross-cut older lateral moraines associated with the Pyke valley and appear undisturbed by more recent glaciations. These offsets draw into question two of Sutherland et al. [2006]'s purported LGM-aged offset hillslopes at the Jerry River. Notice also the redirection of Durwards Creek into the Pyke River between these two times.
assumed that ice retreated from the Alpine Fault at similar times in both locations. A recessional terminal moraine in the lower Kaipo valley ~4 km northwest of the Alpine Fault may be age correlative to the CA5 recessional moraines of *Sutherland et al.* [2007b] in the lower Cascade valley, indicating a post-19 ka age. Radiocarbon dating of this moraine would provide an important upper limit on a strike-slip rate determination.

A prominent dextral offset of a glaciated hillslope (*Pyke Corner*) between Pyke valley and Big Bay was also utilized by *Sutherland et al.* [2006] (Figure 5.9; see also Figure 5.0). The present study determines a best-fit offset of ~1450 m would restore both the northeast and southwest hillslopes (Figure 5.9). Preservation of this offset may indicate this was the most recent time a major glacier flowed from the Pyke valley into Big Bay. The western hillslopes of *Pyke Corner* are more deeply incised by post-glacial creeks than the northern hillslopes, potentially indicating the western hillslope is related to an older (not synchronous) Hollyford/Hokuri Glacier.

Offset lateral and terminal moraines in the vicinity of Hokuri Creek were examined in detail by *Sutherland* [1995], *Sutherland and Norris* [1995] and *Sutherland et al.* [2006]. Detailed stereopair aerial photographic interpretation and limited fieldwork in the present study has led to a reinterpretation of the extent and correlation of these moraines (cf. Figure 3E of *Sutherland et al.*, 2006 to this study’s Figure 5.1C). I interpret a further lateral/terminal moraine beyond their oldest mapped moraine has been offset ~1700 m and assign it a late Early Otira age of c. 58 ka (their estimate of the oldest recessional moraine). It is worth noting this correlation, as do those of *Sutherland et al.* [2006], assumes there was no spillover of the Hokuri Glacier over the north wall of the Hokuri valley, which may not be a valid assumption. This is an area where airborne LiDAR would be well-suited to allow a more accurate interpretation of offset moraines.

The present study introduces the well-preserved paired Low Creek lateral moraines which have been offset ~2900 m from the Jerry River. These provide crucial cross-cutting relationships with older glacial deposits in the Gorge River area which suggest the Low Creek lateral moraines were formed by a Waimea or early Early Otira advance (Figure 5.2B; Figure 5.10). A Waimea correlation is preferred as the following interglacial period provides more time for the lateral moraines to be offset and thus protected from the next ice advance, although an early Early Otira advance is also plausible. A c. 138 ka age is interpreted using the peak glacial signature associated with this glaciation recorded in EPICA Dome C deuterium data [*Jouzel et al.*, 2007]. The regional ~8000 m dextral offset of valleys and glacial deposits
(Figure 5.4) yields the best-constrained rate by combining a well-constrained offset hillslope at the Cascade River and a well-constrained age for the Wolf Formation and Wolf Tablelands lateral moraines.

Although most of the dextral slip-rates presented in this study have large uncertainties, improved dating (radiocarbon, $^{10}$Be, amino acid racemization) and imaging techniques (airborne and ground-based LiDAR) on the features identified may yield some of the best constrained slip-rates on the Alpine Fault over this span of time.

5.6 Geomorphic Consequences of Dextral Slip on the Alpine Fault

Reversals and/or capture of drainages are not an uncommon occurrence in New Zealand. They have been shown to have resulted from sea level changes (e.g., Kaituna River in Marlborough; Craw et al., 2007a), as well as glacial (e.g., Wairau-Clarence rivers in Marlborough; Craw et al., 2008), and tectonic (e.g., Nevis-Mataura rivers in Otago, Craw et al., 2007b) drivers. As discussed in Supplement 5, even the 2000 m elevation Main Divide of the South Island has been breached in the past, with significant catchment areas reversing flow to be captured by the other side of the divide. The present section focuses on Alpine Fault-induced drainage divides, drainage reversals and drainage captures, which are ubiquitous along the western rangefront of the Southern Alps.

The trace of the Alpine Fault commonly coincides with low saddles between major catchments in part because of the weakness of fault-damaged rocks and in part because fault displacement helps trap and localize surface flow to establish drainages along the fault. Such fault-controlled saddles number well over a hundred along the fault; the present study discusses the features south of the Arawhata River (e.g., examples labeled “Drainage saddle” in Figure 5.6). Drainage capture due to dextral offset (i.e., where a PAC plate stream flowing across the fault is displaced from its AUS plate continuation then exploits the next AUS plate drainage to the south) is a relatively common phenomenon on the Alpine Fault that occurs at a rate dependant on slip rate and the spacing between the drainages. Along much of the western Southern Alps rangefront, including in South Westland, the distance between major catchments at the Alpine Fault is typically 10–15 km. At strike-slip rates of 27–30 mm/yr, this indicates major drainages will be dextrally offset or deflected until c. 200–370 kyrs of time has elapsed at which point the PAC plate drainage will be captured by the next AUS plate drainage to the south. This is well shown by the c. 270 kyr ~8 km dextral offset in which all major river valleys have switched back one valley apart from the Cascade River
Examination of the topography suggests that with a further 2 km of offset (c. 65 kyr) the Cascade River will be captured by the Gorge River to the south and the Martyr and Jackson headwaters will become the headwaters for the Lower Cascade valley.

Dextral movement on the Alpine Fault has allowed the Cascade River to capture the Martyr River from the northeast-flowing Jackson River via the youthful Monkey Puzzle Gorge. The Martyr River was formally the headwaters of the Jackson River until about 1.2 km of dextral offset ago (Figure 5.11). Since no other significant tributaries occur on the Pacific Plate in the 7.5 km to the southwest, it seems likely the Monkey Puzzle Gorge has been carved by the Martyr River within the last 50,000 years. Capture of the Martyr River by the Cascade (by which the Jackson River lost roughly one-third of its catchment area), explains why the Jackson River is noticeably underfit to the size of its lower valley.

The Pyke/Hollyford and Smoothwater/Ellery valleys are the only obvious examples of the direction of flow in a drainage completely reversing (Pyke/Hollyford valley and the Dry Awarua is discussed in detail in Supplement 5). The reversal in the Smoothwater/Ellery seems to have been driven primarily by deposition of glacial deposits, but Alpine Fault displacement (particularly AUS plate uplift) must have played a contributing role as the southeastern side of the valley is dominated by glaciated bedrock exposures. Abandonment of Teer Creek by the Laschelles Creek headwaters was glacially-driven as identified by Sutherland et al. [1995]. Generally speaking, the fast tempo of Alpine Fault displacement out-competes glacial and fluvial modification and exerts the greatest impact on drainage evolution adjacent to the fault.

5.7 Discussion

5.7.1 Implications of spatial variation in strike-slip rate

Considering Norris and Cooper [2001]’s average central Alpine Fault dextral slip rate of 28 ± 5 mm/yr, our 29.6 (-2.1/+2.3) mm/yr rate on the southern onshore Alpine Fault, Barnes [2009]’s rate of 27.2 (-3.0/+1.8) mm/yr for offsets between Milford Sound and George Sound (45 km southwest of Milford Sound), and Barnes [2009]’s rate of 31.4 (-3.5/+2.1) mm/yr for offsets further south near Caswell Sound (70 km southwest of Milford Sound), a relatively constant dextral slip rate of c. 28 mm/yr along c. 440 km of the Alpine Fault south of the Hope Fault is plausible (similar to that proposed by Norris and Cooper, 2001). Another possibility is that there is a c. 3 mm/yr southward increase in dextral slip resolved on the Alpine Fault, as suggested by Barnes [2009]. Regardless, our data do not support a significant
Figure 5.11  Martyr River evolution. LINZ topographic map of (A) present day and (B) ~1200 m dextral offset restored highlighting the relatively recent redirection of the Martyr River from the former headwaters of the Jackson River to a tributary of the Cascade River through the youthful Monkey Puzzle Gorge. This is the largest change that has occurred in the Cascade catchment in the last > 300 kyrs.
along-strike deficit in Alpine Fault strike-slip rate in South Westland, despite the structural complexity of the PAC plate here [e.g., Rattenbury et al., 2010; Turnbull, 2000; Chapter 2] including faults which have been reactivated since the last glacial maximum (< 20 ka) [e.g., Sutherland, 1995], and the region hosting the only major change in continental/oceanic lithosphere adjacent to the fault. Such a constant along-strike strike-slip rate is understandable given the relatively smooth trace and uncomplicated nature of the Alpine Fault in the region and an 8000 yr paleoseismic record of large magnitude, surface-rupturing Alpine Fault earthquakes in the region (Hokuri Creek), which documents 24 earthquakes with near-regular recurrence and suggests a tendency for the fault to work in isolation from other faults [Berryman et al., 1992, 2012a]. Single-event horizontal displacements on the Alpine Fault in South Westland have been measured to be c. 7.5–9 m [Sutherland and Norris, 1995; Berryman et al., 2012a, 2012b]. Using a 329 ± 68 year mean recurrence interval determined by Berryman et al. [2012a] from the 8000 yr earthquake record at Hokuri Creek and our slip rate of c. 29.6 mm/yr, single-event horizontal displacements should average c. 10 m (i.e., larger than the last two displacements recorded at Hokuri Creek). This means Alpine Fault single-event horizontal displacements > 10 m could be expected.

A dextral slip rate of 29.6 mm/yr equates to 81% of current AUS-PAC plate boundary motion (83% of Alpine Fault-parallel motion) [DeMets et al., 1994]. The remaining c. 20% of plate motion is accommodated on other low slip-rate structures broadly distributed across the South Island [e.g., Norris et al., 1990; Wallace et al., 2007]. Major faults in the South Westland region that may be presently accommodating plate motion include the Hollyford, Pyke, Glade-Darrans, Pembroke and Moonlight faults, as well as NE- to E-striking structures identified in the Olivine and Skippers ranges and those associated with the offshore north-propagating subduction zone [e.g., Ballard, 1989; Sutherland, 1995; Barnes et al., 2005; Chapter 2]. While the remaining plate motion may be concentrated on Alpine Fault-proximal faults in South Westland (i.e., the deformation is less broadly distributed than to the north), these faults appear to have little effect on regulating the timing of Alpine Fault earthquakes (as suggested by Berryman et al., 2012a) and do not take slip away from the Alpine Fault. Despite high structural complexity in this region, the Alpine Fault is a simple structure that tends to work in isolation from other faults.
5.7.2 Implications of temporal variation in strike-slip rate

The Alpine Fault is a remarkably straight and localized structure. Bedrock exposures of the Alpine Fault zone reveal a 1–12 m-wide zone of principal slip (i.e., fault core) localizing modern slip in the shallow crust and a c. 1 km-wide package of c. 5 Ma exhumed ductile fault rocks (i.e., protomylonites, mylonites and ultramylonites) thought to be comparable with the localized zone of shear presently active at middle to lower crustal levels beneath the seismogenic portion of the Alpine Fault [e.g., Sutherland and Norris, 1995; Batt et al., 2000; Norris and Cooper, 2003; Norris and Cooper, 2007; Chapter 2]. High shear strains (mylonites: 120–200; ultramylonites: 180–300) have been recorded in the exhumed mylonite zone and are consistent with the rate of surface displacement being accommodated in a 1–2 km-wide mylonite zone in the middle and lower crusts [Norris and Cooper, 2003]. Geophysical and geological data are in good agreement with finite element models; both suggest deformation is localized on a narrow ductile creeping zone extending to at least 30 km depth [e.g., Ellis et al., 2006].

Sutherland [1994] and Sutherland et al. [1995] used the reasonably well-constrained age of the Halfway Formation conglomerate (3.6 ± 0.5 Ma) exposed near the mouths of the Cascade and Hope rivers and the clear Fiordland provenance of its clasts (requiring a minimum strike-slip offset of 95–100 km) to determine a minimum Alpine Fault dextral slip rate of ≥ 27 ± 4 mm/yr; this is consistent with our estimated rate over the same region and I suggest the dextral slip rate of the Alpine Fault may have been relatively constant over the last c. 3.5 Ma. Because this weak and highly-localized plate boundary structure does not seem to absorb an ever-increasing share of the total AUS-PAC plate boundary rate with time, and instead accommodates a constant dextral slip rate less than the plate boundary rate, the fault may be considered a mature structure that is already as weak as it can be. Potentially surface displacement on the Alpine Fault may be dictated by rate-limited middle to lower crustal processes of localized ductile creep beneath [e.g., Scholz, 1988]. I speculate ductile processes at middle to lower crustal depths may limit the rate at which the upper crust is able to be loaded, and thus accommodate slip on the fault; this would explain in part why the remaining c. 20% of plate motion is accommodated elsewhere.

5.8 Conclusion

Over the last > 500 kyrs climatic shifts have come and gone and the landscape has progressively evolved, with the Alpine Fault remaining the only constant. Southern Alpine
Fault dextral slip rates presented in this study are relatively constant slip at 27–30 mm/yr over the last > 300 kyrs, and potentially over the last ≥ c. 3.5 Myr, which lead me to suggest the fault is a mature, evolved structure incapable of accommodating more than its present strike-slip rate. The remainder of the c. 20% plate boundary motion not accommodated on the Alpine Fault must then be partitioned on to other structures. These dextral slip rates are consistent with previously determined rates both to the north and to the south indicating a smooth southward increase in Alpine Fault dextral slip rate is plausible, as proposed by Barnes [2009]. Alpine Fault-proximal Australian plate uplift rates of 2.2–2.5 mm/yr have also been relatively constant in South Westland over the last c. 300 kyr. The constancy of Australian plate net uplift and strike-slip rates here provides a rare opportunity to reconstruct paleotopography and determine the timing of glaciations, drainage reversals, and other landscape evolution events not recorded elsewhere in Westland. A remarkable ~8 km dextral offset indicated by reconstruction of geomorphology along ~100 km of the Alpine Fault indicates the Waimaunga Glaciation (MIS 8) was a major formative event in the sculpting and establishment of the modern river valleys here, and presumably elsewhere in the Southern Alps.
Figure 6.0 Oblique aerial photograph looking southwest along the well defined trace of the Alpine Fault from above the Duncan River. Notice the prominent northwest-side-up scarp and small pull-apart basin (same one in the Frontispiece) in the fore- to mid-ground and the lower elevation of the ranges west (right) of the fault despite higher uplift rates. This is a very dynamic landscape dictated by the plate boundary tectonics of the Alpine Fault. On the horizon the Alpine Fault plunges into the Tasman Sea and continues offshore for another ~200 km. The entire field area can be seen between this photo and the Frontispiece. Photo taken by L. Homer (A7742A). Photo courtesy of GNS Science.
6.1 Synthesis

This thesis employed an integrated field-focused approach and applied diverse laboratory analytical and experimental techniques to reveal and highlight new aspects of the tectonophysics, geomorphology and geologic hazard of the central and southern Alpine Fault. Specifically, the goal of this study is to better understand (1) the geometry of the Alpine Fault zone and the physics that govern this, (2) the extent of geological hazards that will be associated with the next Alpine Fault earthquake, which could be \( M_W \geq 8 \) or more, and (3) the tectono-geomorphic history of the Alpine Fault.

During detailed fieldwork I have documented three complete sections through the fault core of the southern Alpine Fault. Observations of the Alpine Fault damage zone and tectonic geomorphology at these locations indicates that modern slip is localized to a single 1 to 12 m-thick fault core. This core is composed of impermeable (\( k = 10^{-20} \) to \( 10^{-22} \) m\(^2\)), frictionally weak (\( \mu_s = 0.12 - 0.37 \)), velocity-strengthening, illite-chlorite and saponite-chlorite-lizardite fault gouges. These phyllosilicate-rich fault gouges are: (1) comparable in structural and mechanical characteristics to those of other major weak-cored faults (e.g., San Andreas Fault), and (2) compatible with fault creep on this section of the Alpine Fault, despite abundant paleoseismic and geomorphic evidence of quasi-periodic large magnitude (\( M_W > 7 \)) earthquakes. I suggest that coseismic slip is localized on the zero thickness fault core margins, while phyllosilicate-rich fault gouge in fault core interiors that exhibit variably oriented, often anastomosing, foliations may accommodate slower rate deformation. The frictionally-weakest fault gouge occurs where the fault core is widest, and is spatially associated with the Kaipo Mélange, a post-Late Oligocene (potentially post- Middle Miocene) serpentinite-bearing tectonic mélange adjacent to the Alpine Fault for \( \sim 40 \) km.

Despite transecting diverse lithologies and intersecting faults, the southern Alpine Fault zone is a relatively uncomplicated and localized structure along much of its length. Comparatively little slip is presently partitioned on to adjacent faults, presumably due to the observed weakness and continuity of the Alpine Fault’s fault core. An abrupt change in along-strike kinematic slip (and uplift polarity) from strike-slip dominant dextral>normal-motion on the southern Alpine Fault and oblique dextral-reverse motion on the central Alpine Fault is constrained to occur at the Martyr River without stepping or significant steepening of the fault plane.
In contrast to the relatively straight and localized dextral-normal-motion fault traces of the southern Alpine Fault, non-optimally-oriented oblique dextral-reverse motion on the central Alpine Fault provides conditions and fault-bounding lithologies favorable for 1–10 km-long serially-partitioned oblique-thrust and strike-slip faults to develop in the upper ~1–2 km. Interpretation of airborne light detection and ranging (LiDAR) data collected for a portion of the central Alpine Fault confirmed that previous mapping of serially-partitioned faults was broadly correct, but revealed the previously-identified structures are made up of multiple fault strands an order of magnitude smaller, and revealed the widespread occurrence of ~300 m-wide parallel-partitioned positive flower structures superimposed on the serial-partitioned faults. A fault kinematic analysis predicts the fault trace orientations observed and supports the concept that the partitioning behavior is scale dependent, with different mechanisms (i.e., crustal-scale discontinuities, thermal weakening, fluvial incision, sediment interaction, damage zone width) exerting control at different scales (< $10^6$ to $10^0$ m). A slip stability analysis suggests that the newly-formed, shallowly-rooted, surface-rupturing faults are kinematically stable, and thus existing fault traces are expected to reanimate. The ~300 m-wide band of fault traces then defines a surface rupture hazard zone where future ruptures are expected to occur, and where habitations and other infrastructure should not be built.

Deep-seated, long runout, catastrophic rock avalanches pose an appreciable and currently underappreciated hazard of Alpine Fault earthquakes. Much of the western rangefront of the Southern Alps has glacially-oversteepend slopes and high relief. With large volumes on the order of ~0.01–1 km$^3$ and long runouts on the order of 5 km, catastrophic rock avalanches are capable of directly or indirectly (through upstream dams and sediment pulses) impacting a large area. Despite its size and remarkable preservation, the ~0.75 km$^3$ Cascade rock avalanche deposit has been previously identified as a terminal moraine and fault-damaged outcrop, highlighting the common misinterpretation of similar rock avalanches in these environments. The Cascade rock avalanche is interpreted to have been coseismically triggered by a large to great Alpine Fault earthquake estimated to have occurred at c. 660 AD, similar to the Round Top rock avalanche south of Hokitika at c. 860 AD. The Cascade rock avalanche is notable for its unambiguous structural relationship to the Alpine Fault, as well as to pre-existing deep-seated bedrock failures, the latter common throughout the Southern Alps. In comparison with other well-documented rock avalanches in the Southern Alps and Fiordland, it is proposed catastrophic failure of pre-existing deep-seated slope failures may be more common than currently documented and characteristics of existing failures may be used to identify locations where a future failure could occur. Of particular note is a mass directly
above the town of Franz Josef, which displays several of the structural characteristics identified at the Cascade and Round Top rock avalanches, and poses a considerable risk to the town below.

Over the accumulation of about a thousand surface-rupturing earthquakes on the southern Alpine Fault, dextral-normal Alpine Fault offset and net uplift of the Australian plate has allowed remarkable preservation of uplifted Quaternary marine sediments to elevations of ~600 m and a ≤ c. 300 kyr record of glacial deposits protected from subsequent glaciations by offset from their source catchment. A remarkable ~8 km dextral offset of major valleys and glacial deposits is recorded along ~100 km of the southern Alpine Fault. Paleontologic and stratigraphic age constraints on c. 300 ka regressive marine sediments yield fault-proximal Australian plate uplift rates of 2.2 (-0.4/+1.2) mm/yr and 2.5 (-0.4/+0.7) mm/yr, and ~8 km offset overlying glacial deposits provide an Alpine Fault dextral slip rate of 29.6 (-2.1/+2.3) mm/yr since their deposition. In conjunction with other dextral rates presented in this study and a re-evaluation of the slip-rate and uplift rate catalog for the southern Alpine Fault, it is suggested these rates have been relatively constant over the last > 300 yrs, and potentially > 3.5 Myr. Australian plate mylonites at the Martyr River (which would have been formed at a location to the southwest) were mylonitized before 70 Ma, indicating that while the southern Alpine Fault follows a pre-existing locus of deep-seated shear, its motion since c. 24 Ma has been insufficient to exhume mylonites on the Australian plate (i.e., its motion has remained strike-slip dominant over this time). In contrast, c. 18 Ma Pacific plate mylonites at the Martyr River (which would have been formed at a location to the northeast) are some of the oldest and southernmost Alpine Fault-related mylonites exhumed by the dextral-reverse central Alpine Fault, which have then been translated to their present position by strike-slip motion. This requires that reverse slip has been accommodated within a localized zone in the central Alpine Fault zone for much longer than suggested in previous studies.

In synthesis, the results of this study present a view of the southern half of the Alpine Fault as a highly-localized, long-lived, very weak locus of plate boundary motion that has had relatively constant spatio-temporal displacement rates in the latter part of its history, ruptures in hazardous large magnitude earthquakes with considerable coseismic geomorphic effects, but may also accommodate some displacement at slower rates, and exerts a first-order control on landscape evolution of the South Island.
6.2 Future Work

Of this study's contributions, not least is the plethora of questions which were introduced but remain unanswered. Addressed here are topics the author considers worthy of further research.

* Recently-acquired high resolution bathymetric data offshore Fiordland has proved very useful in delineating offshore fault traces [e.g., Barnes et al., 2005; Barnes, 2009] and should be extended through Jackson Bay (paired with a seismic survey). The offshore region between Big Bay and Jackson Bay coincides with an important transition between oceanic and continental crust (i.e., the edge of the Challenger Plateau). Fault activity offshore here is a large unknown and an understanding is necessary to constrain plate boundary partitioning and place a northernmost limit on Puysegur subduction-related structures (and associated seismic hazard). It is also likely that a flight of wave-cut terraces (i.e., old beaches) are preserved on the continental shelf here since a ~130 m low-stand about 120,000 years ago (the present study identifies an escarpment in 80 m water depth that may correlate to the -130 m low-stand but better resolution is needed). Correlation of the depth of terraces to a eustatic sea level curve or other marine terrace records (e.g., as Bull and Cooper, 1986 have done for emergent terraces in central Westland) could provide spatio-temporal coastal uplift variations across the region.

* A mélange identified in this study (herein named the Kaipo Mélange) was observed adjacent to the Alpine Fault on the Australian Plate at the Kaipo Slips, east of Lake McKerrow, Hokuri Creek and McKenzie Creek, but likely extends from the Pyke River to offshore John O'Groats (a distance of 40 km) and is deserving of further study. The character and composition of the mélange varies markedly along-strike. Determining the sources and ages of the mélange lithologies (serpentinites, intrusives, limestones, shales), its relationship to the adjacent Western Province rocks, and better defining its deformation style (mechanisms and resultant structures) would potentially allow me to define its significance in the context of the Alpine Fault. For example, is it related to the brittle broken formation overprint on Australian plate mylonites further north at the Martyr River? In particular, attempting to geochemically fingerprint the serpentinites (i.e., whether they are Dun Mountain Ophiolite, Anita Ultramafics, Lone Stag Ultramafics or a foreign Australian plate lithology) would be particularly useful for understanding the mélange's genesis. Chemical analyses of magnetite-rimmed primary chromium spinels may be used to fingerprint the ultramafic source, as demonstrated by Moore and Rymer [2012] for the central San Andreas Fault. This mélange is a...
very important source of weak materials within the Alpine Fault core and thus for local fault mechanics.

* The McKerrow triangulation network across the Alpine Fault was observed in 1976, 1978, 1981 and 1984 and has not been re-occupied since. Repopulating this network (if it is still intact) with a differential GPS or total station survey would extend the range of this dataset by more than 28 years, a time period which should be adequate to record whether a component of fault creep occurs on the Alpine Fault near Hokuri Creek, where the fault core is dominated by very weak material. Maintenance of this network for a future major-great Alpine Fault earthquake could lead to a rich temporal dataset of post-seismic deformation.

* A better understanding of the Pembroke and Glade-Darrans faults is needed; (i) The Pembroke Fault has a length ~40 km, strikes parallel to the Alpine Fault and is separated from it by only 3 km, has a well-defined geomorphic trace that transects low saddles eroded into fault-damaged rocks, has a pronounced regional geomorphic step across it, and has a very large ~1 km$^3$ rock avalanche (John O’Groats rock avalanche) source area immediately above its trace. Little is known about recent activity on the Pembroke Fault, with the exception of a Late Quaternary fault scarp in the Transit River [Turnbull et al., 2010]. (ii) The present study proposed a link between lithologies of the Anita Shear Zone and those of the northwest Skippers Range across the Glade-Darrans Fault- more petrology, geochemistry and radiometric dating is needed to confirm this link, which may provide further constraints on the timing of Fiordland uplift. The Glade-Darrans Fault is also continuous for over 40 km, forms the major geomorphic and lithologic boundary between the Southern Alps and Fiordland, and has well-defined fault traces with fault-damaged rock and clay gouge [e.g., Sutherland, 1995]. Despite its notable geomorphic expression, it is presently unknown if it is a Quaternary-active structure.

* In the author’s opinion the McKenzie Creek area has the best Alpine Fault damage zone and fault core outcrops observed anywhere along the fault. The present study was the first geological study to visit the area and unfortunately only a few days were spent in the area, during which the weather was poor. The area is deserving of further study.

* The Cascade River catchment has the most diverse lithologies of any major catchment in New Zealand and would be well-suited for a detailed provenance study of bedload and comparison to glacial and fluvial deposits to understand how these two processes transport material differently. In particular, the Cascade would be an ideal place to study how
provenance of a sediment downstream of a new source lithology reflects the inputs from that source. This study would also yield data on the relative erosion-resistance of different rock types in this environment.

* Barrier Formation stretched pebble conglomerates are exposed for nearly 60 km from the Alpine Fault near the Cascade River to the Barrier River along the Hollyford Fault System. As originally pointed out by *Sutherland* [1995], across this distance the recorded strain, metamorphic mineral assemblage and provenance varies considerably. A detailed study could provide useful metamorphic and strain gradients along the Hollyford Fault System, one of the major fault systems in the area, which at least locally, has recent surface ruptures, but may have been more active in the past.

* Airborne LiDAR should be flown along as much of the Alpine Fault as possible (but particularly from the Jackson River to the Pyke River) to overcome the difficulties in interpreting fault traces (either aerially or on the ground) in the dense podocarp rainforest. A better understanding of reactivated and new structures on the Pacific Plate adjacent to the Alpine Fault in the Jackson River-Pyke River region is needed. The diverse lithologies and abundance of intersecting faults make this one of the most intriguing regions for future Alpine Fault research.

* Airborne LiDAR should be flown across the Cascade Plateau from Cascade Point to the Alpine Fault to provide better constraints on geomorphic expression of Australian plate deformation. The superb lateral moraine crests here are remarkably continuous linear markers along the 14 km from Cascade Point to Laschelles Creek. The resolution of the LiDAR would likely allow correlation with lateral moraine crests and glacial trim lines that extend all the way to the Alpine Fault west of the Martyr River, to give a 20 km long transect of linear markers across the full width of the sub-aerial Australian Plate. While it is known that nowhere are these glacial features offset by faults, it is unknown exactly how structures resulting from other mechanism of uplift (folding and tilting) may vary across the Australian Plate. This work may have implications for West Coast townships further north. It would also allow work by *Sutherland et al.* [2007b] on the Cascade Plateau to be extended for a better understanding of climatic cycles in New Zealand, and would also be important in terms of developing an integrated Alpine Fault model as the Australian Plate is largely assumed to be a rigid indenter.
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*10Be cosmogenic radionuclide dating on the sackung scarps above the Cascade Rock Avalanche could help determine the relative timing of slip on these topography-rooted failure surfaces. A detailed vertical transect down the sackung scarp correlating to the main failure surface of the Cascade Rock Avalanche could reveal whether the scarp formed all at once or as a series of events through time, a fundamental question to understanding the hazard associated with future deep-seated slope failures here, and elsewhere in New Zealand. The Cascade Rock Avalanche is an ideal candidate for further study because of its relationship to the Alpine Fault, obvious failure surfaces, well-preserved geomorphology, excellent outcrop, and potential for repeated failures in the future (which could dam the river and upset settlement and farming in the lower Cascade valley). More careful radiocarbon dating of entrained beech logs will improve the age control on the Cascade rock avalanche, potentially yielding the best constrained age for the associated Alpine Fault earthquake (as other dates for this event rely on the dating of post-seismic sediment pulses).

* An airborne geophysical survey has recently been flown over the Westland district by the New Zealand Petroleum & Minerals, Ministry of Economic Development, including the northern portion of the study area south to the Pyke River. This survey is slated for public release in early 2013 and should be an important tool for confirming geological interpretation in the area. The aeromagnetic survey data should be particularly useful for delimiting rock avalanche and moraine deposits (which will have a unique signature depending on the lithologies involved) and the position of faults bounding different lithologies in dense bush.

* Further work should be done to date glaciations in southern South Westland. As suggested by this study, the region west of the Alpine Fault from Milford Sound to the Arawhata River contain some of the most extensive and longest-lived records of glaciations onshore New Zealand. More detailed (i.e., quantitative) studies of the clast lithologies within all glacial deposits would contribute to the correlations proposed by this study.

* The fossiliferous marine sand unit identified adjacent to the Alpine Fault on the Australian Plate at McKenzie Creek (Sara Formation) contains remarkably well-preserved and delicate molluscs in an uncemented sand. These shells are highly-fractured and it is obvious the fracturing occurred post-deposition due to the fine preservation otherwise. Due to the proximity of the Alpine Fault (and shallow burial depth), it is possible the fracturing was caused by earthquake waves propagating through the material, though compaction and/or increase in overburden pressure are a possibility as well. A novel study may be able to use these fractured shells to determine parameters of earthquake waves or ground shaking.
(namely magnitude). Similarly, lacustrine sediments exposed in a fault scarp near the confluence of McKay Creek and the Cascade River appear to be cyclically disrupted (folde, or structure-less). This may be a consequence of proximal fault movement in which case further study of the sequence would provide constraints on earthquake ground shaking.

* Apatite and zircon fission track dating should be done on Australian plate Greenland Group lithologies south of Jackson Bay to extend the dataset by previous workers [e.g., Tippett and Kamp, 1993; Seward and Nathan, 1990] further north. The results would provide good long-term uplift rates that could be compared to those determined from < 290 ka marine deposits in this study and recent fault geomorphic offsets to extend the spatio-temporal record.

* The bulging and oversteepened hillslope above Franz Josef should receive further study as this may present a considerable direct hazard to the township. Fieldwork and airborne LiDAR may prove helpful in better defining structures and topography related to this potential catastrophic slope failure. Sophisticated slope stability analyses should be conducted. These should be used to determine peak ground accelerations that may be capable of triggering a catastrophic failure here.
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A specimen of *Talochlamys gemmulata* found within regressive marine silts of the c. 280 ka Wolf Formation at *Madagascar Creek*, ~560 m above modern sea level.
Supplement to Chapter 2: Nature and Timing of Slip Localization on the Southern Alpine Fault, New Zealand

S2.1 $^{40}\text{Ar}/^{39}\text{Ar}$ Radiometric Dating

S2.1.1 Analytical procedures

Samples were prepared for $^{40}\text{Ar}/^{39}\text{Ar}$ dating at the Western AUS Argon Isotope Facility at Curtin University of Technology (Australia), operated by a consortium consisting of Curtin University and the University of Western Australia. Hornblende separates were pre-treated by leaching in dilute HF for one minute, and both hornblende and mica grains were thoroughly rinsed with distilled water in an ultrasonic cleaner. Samples were then loaded into individual wells in an aluminum disc for irradiation, bracketed by Fish Canyon sanidine (FCs) as a neutron fluence monitor, for which an age of 28.03 ± 0.08 Ma was adopted [Jourdan and Renne, 2007]. The discs were Cd-shielded to minimize undesirable nuclear interference reactions and irradiated for 25 hours in the Hamilton McMaster University nuclear reactor (Canada) in position 5C.

Mica samples were analysed by step-heating individual ~0.5 mm diameter grains with a 110 W Spectron Laser Systems continuous Nd-YAG (11064 nm) laser, rastering the beam over the sample for approximately 1 minute for each heating step to ensure homogenous distribution of temperature. Approximately 30 mg of hornblende sample OU82946 was packaged in zero-blank niobium foil and step-heated in a double vacuum high frequency Pond Engineering furnace.

Gas was purified in a stainless steel extraction line using three SAES AP10 getters and a liquid nitrogen condensation trap. Ar isotopes were measured across 10 cycles of peak hopping using a MAP 215-50 mass spectrometer operated in static mode with a Balzers SEV 217 electron multiplier. Data acquisition was performed in a LabView environment with the Argus program written by M. O. McWilliams. Raw data were processed using ArArCALC software [Koppers, 2002], and the ages quoted have been calculated using the decay constants recommended by Steiger and Jäger [1977]. Laser blanks were monitored every 3 to 4 steps and typical $^{40}\text{Ar}$ blanks range from $1 \times 10^{-16}$ to $2 \times 10^{-16}$ mol. Furnace blanks were monitored every 3 samples and range from 3 to 10 times the laser blanks.
S2.1.2 Sample petrology

Sample OU82942 is a weakly mylonitized foliation-parallel quartz-feldspar leucosome hosted in calcic plagioclase (hornfels facies) metamorphic zone Greenland Group metasandstone and metapelite from Spoon Slip, 12 km southwest of the Martyr River, and about 150 m northwest of the Alpine Fault (metamorphic zones after Rattenbury et al., 2010). OU82943 is a sillimanite (amphibolites facies) metamorphic zone Greenland Group gneiss (the highest metamorphic grade this unit reaches) from Monkey Puzzle Gorge, about 1 km northwest of the Alpine Fault at the Martyr River. OU82944 is a low grade quartzo-pelitic mylonite with anatomosing foliation and a semi-brittle overprint recovered about 90 m northwest of the Alpine Fault in a tributary of the Martyr River. OU82945 is a well-foliated quartzo-pelitic mylonite from 50 m northwest of the Alpine Fault at the Martyr River. Outcrops clearly show the low grade mylonitization in OU82944 and OU82945 overprints the sillimanite grade gneissic fabric in OU82943. OU82946 is a rare 2m by 5 m lozenge of L > S fabric hornblende dioritic mylonite hosted within a 400 m wide section of fine-grained chlorite-epidote-quartz metavolcaniclastic mylonite on the PAC plate. Thin section observations indicate the chlorite-epidote alteration occurred at a similar time to the mylonitization because these minerals are drawn out parallel to the mylonitic foliation (and likely at the same time as the host Brook Street terrane-derived volcanioclastic mylonites). Ductile folding of both AUS and PAC mylonites appears to post-date shear fabrics as these fabrics are folded. Detailed petrology of these samples is presented in the Appendix.

S2.1.3 Expanded results

Results from $^{40}$Ar-$^{39}$Ar analyses are summarised in Figure 2.7A as age spectra plotting variation in the nominal age for each heating step against progressive release of $^{39}$Ar as a proxy for experimental progress towards complete outgassing. Additional isotopic characterisation is provided for hornblende sample OU82946 by an inverse isochron plot ($^{39}$Ar/$^{40}$Ar ratio against $^{36}$Ar/$^{40}$Ar ratio) (Figure 2.7B). Such interrogation was not considered relevant for the mica samples, as data from these tend to cluster near the radiogenic axis and exhibit a poor spread along nominal mixing lines, preventing the robust distinction of isotopic trends.

With the possible exception of OU82943, the age spectra express a common general form, with young initial ages increasing to older levels over the bulk of gas release defining an irregular plateau (mica OU82945 and hornblende OU82946) or exhibiting mild dispersal
about a general attractor (OU82944 and OU82942). Such patterns are typical of partially disturbed systems, where a thermal and/or mineralogical overprint has reset the argon systematics of exterior and less retentive sites within the crystal lattice. In such systems, significant geological meaning can be derived from the young initial ages – which date the nominal episode of disturbance – and to a lesser degree, the older age attractor. Although rendered imprecise by diffusional mobility of $^{40}$Ar during the disturbance episode, the older age attractor provides at least a qualitative guide to the inherited age on which the disturbance was imposed.

This breakdown is best defined for hornblende sample OU82946, for which the initial lower age defines an extended plateau rather than a simple asymptotic limit. $^{39}$Ar/$^{40}$Ar-$^{36}$Ar/$^{40}$Ar isotope ratio trends demonstrate equilibration of this sample with non-atmospheric argon reservoirs, such that elementary correction for non-radiogenic argon mis-estimates sample ages. Fitting reverse isochrons through relevant data points to more appropriately correct for this mixing produces model ages of 18 ± 4 Ma for the initial plateau and 294 ± 10 Ma for the older heating steps later in the experiment.

Although less precisely defined and not individually compelling, the initial disturbance ages for mica samples OU82945, OU82944 and OU82942 are all consistent with the Early Miocene episode postulated for OU82946, supporting the attribution of wider geological significance to this age. The older ages for these samples are more varied, with imprecisely defined attractors at approximately 260 Ma for OU82942 and 210 Ma for OU82944, and a rigorous plateau age (comprising sequential steps with statistically indistinguishable apparent ages) of 142.9 ± 2.4 Ma across the last 47% of argon release for sample OU82945. Although limited to a largely qualitative definition, the two-fold variation in these older signals indicates that this probably reflects real variation in the pre-Miocene fabric ages of these samples.

Mica sample OU82943 departs from this general behavior, with progressive age release dominated by mild dispersal about an attractor at c. 235 Ma throughout the bulk of gas release. Although exhibiting slightly reduced and variable initial ages, no clear evidence of a specific disturbance episode can be resolved for this sample.

### S2.2 Crystallographic Preferred Orientations of PAC and AUS plate mylonites

A pilot study of 12 polished thin sections of mylonites and gneisses from near the Martyr River were analyzed using electron backscatter diffraction (EBSD) to collect crystallographic
preferred orientations (CPOs). From these, I determined quartz slip systems active during
deformation, and assessed CPO strength to compliment $^{40}\text{Ar}/^{39}\text{Ar}$ dating on comparable
samples. This was done using standard EBSD techniques [e.g., Prior *et al.*, 1999; Barth, 2008
and references therein] on a FEI Quanta 400F field emission source SEM equipped with an
HKL Technology Nordlys 2 detector at the University of California- Santa Barbara.

Samples chosen for this study are polycrystalline, but have high percentages and high
interconnectivity of quartz. Samples were oriented in the field and cut perpendicular to
foliation and parallel to the lineation (X-Z sections; cut along a 070 direction if lineation was
not present). All the data relating to these samples is available on the DVD. Presented here
are the 4 samples most pertinent to the $^{40}\text{Ar}/^{39}\text{Ar}$ study (Figure S2.1).

Interpretations of CPOs are after Figures 4.40–4.44 of *Passhier and Trouw* [2005] and Figure 1
of *Barth et al.* [2010]. OU82963 is of a foliation-parallel quartz-rich horizon in a PAC plate
chlorite-epidote volcaniclastic mylonite with a Brook Street terrane protolith (similar to
sample OU82946); the quartz CPO is strong and indicative of dextral non-coaxial flattening
at low temperatures (e.g., greenschist facies) accommodated by basal $<a>$ slip. OU829431 is
a PAC plate sillimanite-grade quartzo-pelitic Greenland Group gneiss from Monkey Puzzle
Gorge (similar to OU82943); it has a weak quartz CPO consistent with the annealed quartz
microstructures observed in OU82943 and OU829431 (see Appendix). OU82960 is a low
grade AUS plate quartzo-pelitic Greenland Group mylonite (similar to OU82944) from ~100
m horizontal distance from the Alpine Fault; it has a strong Type I crossed girdle CPO
indicative of coaxial plane strain deformation at low grade metamorphic conditions (basal
$<a>$ slip and rhomb $<a>$ slip). OU82964 is an AUS plate quartzo-pelitic Greenland Group
mylonite from ~20 m from the Alpine Fault (similar to OU82945); it has a moderately strong
CPO indicative of sinistral non-coaxial flattening at low temperatures accommodated by basal
$<a>$ slip.

**S2.3  Carbonate C and O Stable Isotopes**

The excellent outcrop on both sides of the Alpine Fault core at the Martyr River and
tributaries of McKenzie Creek presents a rare opportunity to address a fundamental question
regarding the role that fluids play during seismic cycling on major fault; namely, does seismic
rupture temporarily allow fluids to communicate through a fault core that has low
permeability in the interseismic period? On the San Andreas Fault, for example, it has been
found the low permeabilities of phyllosilicate-rich fault core rocks isolate the fluid regimes on
Figure S2.1  Select pole figures of quartz grain orientations in samples as indicated. Pole figures are one point per grain and utilize a lower hemisphere stereographic projection. Pole figures for \{0001\} (c axis) and \{1-100\} and \{11-20\} (a axes) are shown. See text for comparison to \(^{40}\)Ar/\(^{39}\)Ar dated samples.
the PAC and North American plates throughout the seismic cycle [e.g., Carpenter et al., 2011; Wiersberg and Erzinger, 2011]. At Martyr River and McKenzie Creek localities, the fault core consists of a throughgoing > 1.5 m thick zone of phyllosilicate-rich gouge with experimentally-measured permeabilities of $10^{-20}$ to $10^{-22}$ m$^2$ (Chapter 2) and thus the fault core should act as an impermeable seal, isolating the hydrologic regimes in the hangingwall and footwall. To address this question, a pilot study of 10 calcite vein samples were collected from veins cross-cutting fault rocks within 25 m either side of the fault core for $\delta^{13}$C and $\delta^{18}$O stable isotope analysis.

### S2.3.1 Analytical procedures

Calcite samples were crushed to a fine to very fine grain size using a mortar and pestle. Pristine grains were handpicked using a Nikon SMZ-U binocular microscope. Each sample was then weighed into two duplicate sub-samples weighing exactly 100 µg using a Mettler AT261 DeltaRange scale. Both sub-samples of each sample name were analyzed for a total of twenty analyses.

Samples were prepared and analyzed for $\delta^{13}$C and $\delta^{18}$O stable isotopes by Isotrace Research (Department of Chemistry, University of Otago) utilizing the phosphoric acid dissolution method. The twenty 100 µg samples were reacted with anhydrous phosphoric acid to produce carbon dioxide, using a Thermo Finnegan GasBench. Accurate temperature-control of the reaction conditions was maintained to produce highest precision results. The GasBench sub-sampled the carbon dioxide produced six to ten times for analysis with the Thermo Finnegan Delta+ Advantage isotope ratio mass spectrometer. Samples were analyzed concurrently with the laboratory reference carbonate. Isotopic and elemental values were determined using LSVEC, IRU-Marble, IRU-Bicarb as references. Precision and accuracy were determined using IRU-Marble.

### S2.3.2 Sample petrology

At all outcrops within ~50 m of the Alpine Fault fault core, calcite veins are ubiquitous, but decrease in abundance away from the fault core. All samples collected were of late-stage, hydrologically-precipitated calcite cross-cutting all fault rock fabrics within 25 m (but most samples within 1 m) either side of the low permeability phyllosilicate-rich fault core delineating AUS and PAC plate lithologies. The samples were collected from a few distinct structural contexts including veins, “clots” and “lumps”. Mineralized extensional veins were
the most common sample observed and collected. Calcite in these veins is mostly crystalline and displays textures suggesting single event growth of calcite into the voids. Some veins consist of very fine-grained aggregates of quartz + calcite. Although no cross-cutting relationships were observed between these and other calcite veins, the aggregate quartz + calcite veins can be traced for shorter distances and are therefore presumed to be older; presumably greater vein lengths were truncated during shear on the fault rock fabrics. There are no consistent sets of orientations in any of the veins. These veins are commonly 2–3 mm thick, but occasionally can be greater than 1 cm thick; they are commonly traceable through outcrops for a distance of ~20 cm. “Clots” are formed at the junction of two or more fractures or veins. The intersecting discontinuities may be mineralized to mm-width; alternatively they are hairline-thin and unmineralized. These calcite “clots” are 1–6 cm across and are always coarsely crystalline with cleavage visible in hand sample. “Lumps” refer to masses of calcite ± quartz that occur within cataclastically deformed rocks as isolated clasts or as a part of a clast otherwise containing fault rock. These typically reach a maximum dimension of ~2 cm. Lumps were generally avoided for analysis because of their less certain structural context, but were found to have similar isotopic properties as clots and veins nearby.

S2.3.3 Results

Results of this study are presented in Figure S2.2. Samples from the PAC plate at the Martyr River form a cluster distinct from the AUS plate Martyr River samples with an appreciable difference in δ^{13}C values and overlapping δ^{18}O values. McKenzie Creek samples have less well-defined clusters, but also show a different δ^{13}C values between the AUS and PAC plate-derived samples. When this study’s results are considered in the context of stable isotope results of many different carbonate vein types in the Southern Alps [C.D. Menzies unpublished data, 2012], the results from the Martyr and McKenzie Creek areas are found to be most similar to carbonate that coats joints at Gaunt Creek in the immediate hangingwall of the Alpine Fault (Figure S2.3). These joint-coating calcite veins are the latest stage carbonate precipitation seen at Gaunt Creek. Back-calculating carbonate vein stable isotope measurements reveal that Gaunt Creek joint coatings plot within the range of meteoric-dominated low temperatures veins [C.D. Menzies unpublished data, 2012]; by analogy so do the samples in the present study (Figure S2.4).

Taken together, the late-stage fault rock-hosted carbonate samples analyzed in this study are
interpreted to have precipitated from a low temperature, primarily meteoric-dominated fluid, but with a component of fluid-rock interaction unique to the local lithology on the corresponding side of the fault core. At all localities in this study, the higher $\delta^{13}C$ values are derived from samples on the side of the fault with net modern uplift and the most hydrothermally-altered rocks. The consistent separation of $\delta^{13}C$ and $\delta^{18}O$ values between the two sides of the fault core, particularly at the Martyr, strongly suggest the fault acts as a barrier to hydrologic flow, even coseismically and post-seismically when fault damage creates maximum permeability. Thus, meteoric groundwater infiltrates both sides of the fault core, interacts with the host rocks to a degree, and has no means of communicating with the other side of the fault core. This result complements Sr isotope work [C.D. Menzies unpublished data, 2012] and across-fault fluid pressure and permeability determinations in the DFDP borehole [Sutherland et al., 2012], both on the central Alpine Fault.
S2.4 Geomorphic Constraints for Large Magnitude Alpine Fault Events

The combined dataset of striations on fault planes, juxtaposition of sediments, fault-related damage and meso-scale cataclastic textures, broken trees, proximal rock avalanches, earthquake-correlated sediment pulses in beach dune ridges, and offset surfaces indicate surface displacements that are inferred to have resulted from large magnitude earthquake ruptures that extended to the surface, must occur along the southern Alpine Fault. Offset features (surfaces, drainages, ridges, moraines) indicating dextral and vertical displacements

Figure S2.3  Comparison of stable isotopic data from calcite veins collected in this study to that from others collected elsewhere along the Southern Alps and Alpine Fault [unpublished data courtesy C.D. Menzies 2012]. Abbreviations used: Wkp (Waikukupa Thrust, Alpine Fault), HM (Hare Mare Creek, Alpine Fault), GC (Gaunt Creek, Alpine Fault).
Figure S2.4  Original fluid compositions calculated from carbonate vein stable isotope measurements in Figure S2.3 [C.D. Menzies unpublished data, 2012]. Note that Gaunt Creek joint coatings (most similar to samples analyzed in this study) plot within the range of meteoric-dominated low temperature veins. See Figure S2.3 for abbreviations used.

were identified in aerial photography and in the field. Scarps are typically 2–20 m in height and consistently upthrown to the northwest. Features that preserve both reliable strike-slip and dip-slip offset are rare. Berryman et al. [2012a] note a 7.5 ± 0.5 m dextral offset with about 1 m of up to the northwest motion that they attribute to a single event on the fault near Hokuri Creek. In the adjacent paleoseismic record, a complete sediment cycle associated with one earthquake is ~1 m thick, corresponding to the typical depth of ponding against the fault scarp by coseismic displacement [Berryman et al., 2012a]. Similarly, at all major outcrops identified in this study, Late Quaternary sediments were mapped juxtaposed against at least
one margin of the fault core. At all locations south of the Martyr River, the sediment/fault core interface has shallowly southwest plunging slickenlines and several meters of sediment indicating at least \( \sim 20 \) m of displacement since deposition of the sediments.

Just north of Milford Sound, where the Alpine Fault strikes offshore, evidence for 2 earthquake events is recorded in a trenched fault scarp at the John O’Groats River [Cooper and Norris, 1990]. There, many large trees on the fault scarp have been broken \( \sim 10 \) m off the ground, and have a new branching crown developing from the break [Cooper and Norris, 1990]. Similarly broken trees were observed on the fault trace at McKenzie Creek in the present study. Slope failures of several varieties have been observed along the central and southern Alpine Fault. These are inferred to relate to coseismic fault damage, including active gully/slip systems (Kaipo Slips), inactive gully/slip systems (Saddle Creek area of the Cascade River), landslides (Cascade River) [Korup, 2004] and rock avalanches (e.g., John O’Groats, Pyke, Hope Blue River Range, Cascade rock avalanches; see Chapter 4 and Supplement 4). Whether triggered coseismically or otherwise, these failures deliver large volume sediment pulses downstream [Korup et al., 2004]. A coastal dune ridge at the mouth of the Hollyford Valley has been correlated to a regional episode of progradational dune formation resulting from an earthquake-induced sediment pulse, following the AD 1717 earthquake [Wells and Goff, 2007]. Outcrops of Late Quaternary lacustrine silts on the upthrown side of Alpine Fault scarps in the Cascade and Martyr Rivers have chaotically folded or disrupted laminae that may form in response to coseismic ground motion or soft sediment shearing of the sequence following rapid coseismic uplift; these same silts are comparatively undeformed with distance from the fault scarp. Although geomorphic evidence of fault creep is unlikely to be detected without repeated detailed observations in this environment, all geomorphic evidence points to large magnitude surface-rupturing earthquakes in South Westland. This is found to be in dramatic contrast to the behavior expected for observed fault core gouges, which have mineralogical, microstructural and frictional properties indicative of slow slip rate, probably aseismic processes (Chapter 2).

S2.5 Alpine Fault-Generated Tensile Cracks

Remarkable damage zone tensile fracture sets are exposed adjacent to the Alpine Fault principal slip zone (PSZ) in the hangingwall at McKenzie Creek. Here, PAC plate isotropic metamorphic rocks are pervasively dissected by fractures into \( < 1–2 \) cm-sized cubes within 2 m of the PSZ (Figure S2.5A). Beyond this, over 7 m-width there are very regularly (4–10 cm) spaced fractures in two well-developed tensile fracture sets: the dominant set parallels the
fault core, while the subsidiary set is perpendicular to the first set and perpendicular to the striations defining a slip vector on the fault core margin (Figure S2.5B–C). These same tensile fracture sets are also observed in the cemented ultracataclasite at the Martyr River, but not in foliated rock at a similar distance from the fault plane, suggesting they only form in an isotropic material. These fracture sets are everywhere unmineralized, and their spatial and geometrical relationship to the fault core and slip vector is ubiquitous.

The observations of late-stage tensile cracks and associated pulverized rock suggest a genetic relationship between these structures and the fault core (Figure S2.5). However, current models of tensile crack formation associated with dynamic rupture propagation cannot explain (1) the occurrence of two sets of tensile cracks orthogonal to each other (one parallel to the fault core, the other perpendicular to the slickenlines — i.e., slip vector — on the fault core margin) [e.g., Di Toro et al., 2005; Griffith et al., 2009; Ngo et al., 2012] and (2) the extent of the tensile crack sets, which occur up to 9 m from the fault core. Models of large instantaneous stresses associated with dynamic rupture propagation require supershear rupture velocities to create damage this far from the rupture tip [Yuan et al., 2011]. These structures only seem to form in otherwise cohesive isotropic material adjacent to the fault core, which could explain why they have not been widely reported from other fault zones, which are commonly hosted in anisotropic materials. However, in the absence of other mechanisms for their formation, I suggest that the tensile cracks may be similar to fault-generated pulverized rock [e.g., Dor et al., 2006; Mitchell et al., 2011], albeit formed in isotropic rock.

Fracture densities, fault planes and foliations were examined in detail in a superb outcrop of AUS plate foliated cataclasite located 1.5–4.0 m from the principal slip surface (i.e., immediately adjacent to the phyllosilicate-rich fault core) at the Martyr River (Figure S2.6; see Figure 2.3B for context). Here an S-C foliated cataclasite fabric overprints planar quartzofeldspathic mylonite. The S fabric is defined by Alpine Fault-parallel phyllosilicate-rich fault surfaces, while the intervening C fabric is defined by the long axis of more resistant (quartz-rich) lozenges largely coincident with pre-existing mylonitic fabric. Alpine Fault-parallel fault densities and foliation intensity decrease with distance from the fault core. Length and intensity of tensile fractures also increases away from the fault core here, largely as a function of an increasing abundance and size of more intact mylonite lozenges, within which these features are hosted. Abundance of clay minerals decreases away from the fault core.
Figure S2.5  Brittle Structures. (A) Alpine Fault zone exposure at a tributary of McKenzie Creek showing fault-related tensile fracture sets in an otherwise isotropic PAC Plate lithology. These fractures are confined to a ~9 m-wide zone adjacent to the AUS Plate-derived fault core. Beyond this the fractures observed do not exhibit such preferred orientations. Note that the scale varies in this view. (B) Stereonet of poles and great circles to average orientations of fracture sets shown in A and associated fault core boundary kinematics. Black triangle is the AUS-PAC plate motion vector derived from NUVEL-1A [DeMets et al., 1994]. Black circle is the pole to the dominant fault perpendicular fracture set at a tributary of the Martyr River. Blue square is the slickenline orientation on the fault plane shown in C. Refer to A for other symbols. (C) Slickenlines on the boundary between the saponite-rich AUS Plate-derived fault core and PAC Plate lithologies (Quaternary gravels shown here). Both boundaries of the fault core have slickenlines and offset sediments indicating slip was localized to these two surfaces (see Figure 2.3C and D). This photograph was taken ~100 m along strike of the image in A.
Orthorectified composite photograph of AUS Plate fold样式 cataclasite outcrop 1.5 m from the principal slip surface of the Alpine Fault at the Martyr River (see Figure 2.3 for context). Kinematics are consistent with dextral shear on the Alpine Fault. Tensile fracture density increases away from the fault plane due to the decreasing clay content and increasing size and abundance of more competent lozenges (lithons).
S2.6 Seismicity

The region from Haupiri on the North Westland section to ~20 km southwest of Haast has long been recognized to experience relatively few moderate-sized earthquakes [Evason, 1971; GeoNet, 2012]. However, southwest of this seismic gap, in the vicinity of Jackson Bay and Big Bay, patterns of particularly diffuse seismicity (including strike-slip events) have been identified southeast and northwest of the Alpine Fault [e.g., Scholz et al., 1973; Anderson and Webb, 1994; Eberhart-Phillips, 1995]. Two earthquakes with moment magnitudes (Mw) greater than 5.4 have been centered near the southern onshore section of the Alpine Fault. The 2001 Mw 5.8 Jackson Bay earthquake was a shallow (~4 km focal depth) event located 3–10 km southeast of the Alpine Fault with a focal mechanism consistent with slip on an 048°-striking reverse fault. It also had an aftershock sequence with predominantly strike-slip mechanisms [McGinty et al., 2005]. A M6.2 oblique-reverse event in 1947 had an epicenter 20 km southeast of the Alpine Fault and exhibited an oblique-reverse mechanism [Doser et al., 1999]. South of Jackson Bay, shallow earthquakes (< 15 km depth) become common offshore, correlating with the northern extent of the subducting AUS plate [Doser et al., 1999].

Focal mechanisms of 15 shallow (< 15 km) Mw 4–6 events between Haast and Milford Sound and within 30 km either side of the surface trace of the Alpine Fault yield strike-slip, reverse, normal and oblique focal mechanisms with a relatively consistent P axis of ~122° with a standard deviation of 13° (Figure S2.7) [J. Ristau pers comm., 2012]. Within this region, microearthquakes clustered in time and space do not share a common focal mechanism [Caldwell and Frohlich, 1975; McGinty et al., 2005]. This focal mechanism diversity likely reflects the fact that slip could occur either on inherited or newly-formed structures within the diverse range of lithologies present, and the location at the transition from a dominantly strike-slip to subduction plate boundary [e.g., Doser et al., 1999; Wallace et al., 2007].
Figure S2.7  Centroid Moment Tensor (CMT) solutions for 16 well-located M > 4.0 earthquakes between Milford Sound (south) and the Haast River (north) [data courtesy J. Ristau, 2012]. Latitude and longitude are in decimal degrees. Notice the wide assortment of focal mechanisms across the region and the relatively consistent P-axes trending ~122°.

S2.7 McKerrow Triangulation

Despite evidence for large ground displacements that could occur coseismically [Berryman et al., 2012a], my observations at Hokuri Creek of a 12 m-wide fault core of saponite-dominated foliated fault gouge with a low steady state coefficient of friction of 0.12 and strong velocity-weakening behavior suggest fault creep may also occur [Marone, 1998; Scholz, 2002]. Unfortunately, the depth to which the weak saponite-dominated fault gouge extends is unknown and previous campaign GPS data [Beavan and Haines, 2001] are too sparse in this area to resolve if fault creep has occurred [J. Beavan pers comm., 2010].

A 6 station triangulation and trilateration network spanning the Alpine Fault at Lake McKerrow (near Hokuri Creek) was established in 1976 and was reoccupied repeatedly between 1976 and 1984 (Figure S2.8). Over this time period no measurable component of
fault creep above an uncertainty of ~2 mm/yr was detected (Figure S2.9) [J. Beavan pers comm., 2010]. However, if this network is still intact, reoccupying the stations utilizing a total station GPS and extending this record by another c. 25 years would provide considerably more compelling evidence for, or against, fault creep.

Figure S2.8  Aerial photograph showing the location of the 6 stations used in the triangulation/trilateration network occupied between 1976 and 1984 across the Alpine Fault at Lake McKerrow.
Figure S2.9  Solution to the triangulation network at Lake McKerrow between surveys in 1976 and 1984 (see Figure S2.8 for locations of stations) [data courtesy J. Beavan, 2010]. Vectors and associated error ellipses indicate no component of measureable fault creep (above an uncertainty of ~2 mm/yr) occurred during this time period. Measureable dextral-sense fault creep would cause stations A and X to have shifted consistently northwest (several millimeters or more) relative to stations B, C, Y, and Z.

S2.8  InSAR

Interferometric synthetic aperture radar (InSAR) is an increasingly commonly used technique to determine deformation of the Earth’s surface at centimeter-scale resolution. The technique relies on stacking synthetic aperture radar collected at different times by satellites like ENVISAT; the resulting interferogram can be used to measure geologic spatio-temporal deformation occurring over days (e.g., surface-rupturing earthquakes; Bürgmann et al., 2002;
A pilot study was carried out to test the hypothesis that fault creep may be operative on the southern Alpine Fault in the vicinity of low friction, velocity-weakening fault gouge at Hokuri and McKenzie Creeks. A total of 15 scenes from Track 180 of ENVISAT’s ASAR band spanning 8 years time were ordered from the European Space Agency (ESA). From these data 5 interferograms were produced following the techniques outlined in Ryder and Bürgmann [2008]. Coherence was found to be poor over much of the region due to atmospheric effects and the dense vegetation present. Coherence was best in less vegetated areas such as glacial moraines and alpine areas. The best interferogram produced (Figure S2.10) shows poor coherence along the Alpine Fault in the Hokuri Creek area, so no component of fault creep was detected. However, an interesting result is a +3 mm/yr line-of-sight anomaly spatially coincident with Dun Mountain Ultramafic Group rocks in the Olivine Range near Woodhen and McKay creeks, adjacent to the sinuous trace of the Livingstone Fault Zone. The implication is that these serpentinites are rising (perhaps diapirically?) adjacent to the mélange zone of the Livingstone Fault. A similar anomaly was detected by Ryder and Bürgmann [2008] associated with the New Idria serpentinite in central California (see their Figure 2). The InSAR study yielded a null result with respect to fault creep due to the hindering effects of dense forest cover of the target area and unknown effects of atmosphere variations. More elaborate methods or higher resolution, longer time span satellite data may be able to allow correction for atmosphere variations and produce a more useable interferogram.

S2.9 Potential Diapiric Rise of Serpentinite in Fault Core

Buoyant rise of low-density serpentinite from a deeper source in conjunction with horizontal displacement has been proposed as a possible explanation for the presence of serpentinite within the actively creeping fault cores at SAFOD [Moore and Rymer, 2012]. Such a process could explain (1) the close proximity of fault core hosted serpentinite blocks at Hokuri and McKenzie Creeks to their source despite significant horizontal displacement and slip on both sides of the fault core, and (2) the fact that serpentinite blocks within the fault core are often significantly larger than those found in the adjacent mélange (suggesting they may be sampled from a structural level where the block-in-matrix fabric is coarser). This would imply diapiric rise of the serpentinite blocks occurred at a higher rate than horizontal fault strike-slip rates. Further study is needed to confirm whether this process is operative at Hokuri and McKenzie Creeks.
Figure S2.10  InSAR interferogram produced from scenes of Track 180 of ENVISAT’s ASAR band over 8 years overlain on northwest-illuminated hillshade derived from 25 m LINZ DEM data. Color scale indicates line-of-sight velocities between the satellite and the ground surface (positive is towards satellite, negative is away) in mm/yr. Notice the poor coherence in the forested lowland valleys and saddles along the Alpine Fault. The prominent ~+3 mm/yr anomaly (red) correlates to ultramafic rocks of the Dun Mountain Ultramafic Group in the Olivine Range. The small ~+2.5 mm/yr anomaly further to the north appears to correlate to an area of slope instability (e.g., sackungen) and may be indicative of slope collapse.

S2.10 The Fiordland/Southern Alps/Alpine Fault Junction

The junction between the Alpine, Pembroke, and Glade-Darrans faults at the Hollyford Valley is one of the most significant fault intersections along the Alpine Fault (Figure S2.11). It separates the Southern Alps in the northeast from Fiordland to the southwest, as well as the Anita Shear Zone (between the Pembroke and Alpine Faults), which has distinctly lower relief and lower slopes than neighboring Fiordland. Because the crucial junction between all these faults is beneath Lake McKerrow and because of the varying spatial extents and focuses of previous studies, the geology has been mapped in several different ways (i.e., Pembroke Fault cutting the Glade-Darrans Fault and vice versa, correlation vs. no correlation of units across the Hollyford valley) [Clark and Wellman, 1959; Wood, 1962; NZGS, 1972; Blattner, 1978; Ballard, 1989; Turnbull, 2000]. As pointed out by Turnbull [2000], many of the ages and relationships of the rock units in the area remain unresolved (see also his discussion of these units). Since the present study is one of the few to examine rocks from all of these fault-
bounded blocks, and because of the significance of this junction, a few observations and ideas are summarized here.

Rocks in the Anita Shear Zone (St Anne Formation/Thurso Formation) are lithologically similar to those of the Mt Webb Gneiss and Slabby Peak Schist in the vicinity of Hokuri Creek, although from textural and metamorphic facies viewpoints the rocks are quite distinct (Figure S2.11). Rocks of the 5 km-wide Anita Shear Zone are intensely mylonitized and highly deformed throughout. This mylonitic foliation is deformed by upright isoclinal structures, so the remnant mylonitic shear sense switches from dextral to sinistral (and vice versa) in outcrop over tens of meters. The protolith age is unknown, but the shear zone has been interpreted as a shallow-dipping mid-Cretaceous extensional structure that has been further deformed and metamorphosed in the Cenozoic [Hill, 1995; Klepeis et al., 1999]. By comparison, the Mt Webb Gneiss and Slabby Peak Schist on the east side of the Hollyford Valley are only locally mylonitized. Boudinaged marble lenses in the vicinity of Hokuri Creek are similar to those in the Thurso Formation of the Anita Shear Zone [I. Turnbull pers comm., 2012]. Ballard’s [1989] arguments that the Mt Webb Gneiss is distinct from the Thurso and St Anne formations are unsubstantiated and his definition of Skippers Group is unclear. Mt Webb Gneiss/Slip Hill Intrusives, Hokuri Intrusives, Lone Stag Ultramafics and Slabby Peak Schist are all fault-bounded and distinct from the Twin Lakes Trondhjemite and Mantle Volcanics of definite Brook Street terrane attribution [Ballard, 1989].

The following geologic history is suggested based on observations in this study and those made previously:

(1) The Pembroke Fault or a correlative once extended across the position of the (at that time future) Glade-Darrans Fault to separate the Lone Stag Ultramafics from Brook Street terrane rocks. Mt Webb Gneiss, Slabby Peak Schist, Slip Hill Intrusives, Hokuri Intrusives, and Lone Stag Ultramafics are correlatives of the St Anne Formation, Jagged Gneiss, Thurso Formation, Anita Ultramafics and Milford Orthogneiss.

(2) The Glade-Darrans Fault formed to accommodate northeast-directed thrusting of the Fiordland/Anita block over terranes to the northeast. This fault cut across the Pembroke Fault and exhumed the northern portion of Fiordland sufficiently to juxtapose amphibolites facies strongly-deformed mylonites of the Anita Shear Zone against less-deformed greenschist facies rocks of the northern Skippers
Range. Related compression across this region may have resulted in the closure of narrow fault-bounded Tertiary basins in the region.

(3) Little is known about the Pembroke Fault and Glade-Darrans Faults, but their well-defined traces suggest they may still be active structures.

**Figure S2.11** The Fiordland/Southern Alps/Alpine Fault junction (A) Simplified basement lithology map of the study area showing major faults adjacent to the Alpine Fault overlain on a northwest-illuminated hillshade derived from 15 m LINZ DEM data. Mapping after Rattenbury et al. [2010], Turnbull et al. [2010], Turnbull [2000], and this study. (B) Simplified metamorphic facies map of the region shown in A after Turnbull [2000]. It is suggested the Mt Webb Gneiss, Slabby Peak Schist, Slip Hill Intrusives, Hokuri Intrusives and Lone Stag Ultramafics are lithologically related to rocks units on the Fiordland block (especially the Anita Shear Zone). Subsequently, exhumation of the Fiordland block along the Glade-Darrans Fault juxtaposed distinct mylonitic and metamorphic textures and facies across the Hollyford valley indicating different structural levels.
S3.1 Application of LiDAR to Fault Mapping in Westland Temperate Rainforest

This study has shown that airborne LiDAR can successfully be used to map fault traces beneath dense lowland temperate rainforest in Westland. A 34 km-long by 1.5 km-wide zig-zag flight path was flown to maximize fault coverage (based on previous mapping which indicated the fault is segmented in this way), while minimizing cost (Figure S3.1). Data were collected in six overlapping swaths ~780 m-wide (for a total width of ~1.5 km) to increase the point counts through the dense podocarp rainforest; ground returns through the rainforest canopy and understory often averaged only about 1 point per 1 m², which was sufficient to produce a 2 m digital elevation model (DEM). The result is an unprecedented view of the bare-earth landforms beneath the vegetation and over an order of magnitude resolution improvement over existing landscape models (Figure S3.2). Figure S3.3 highlights the difficulties in observing fault geomorphic features in the dense podocarp rainforest. A paper detailing the methods used in the central Alpine Fault LiDAR survey (and surveys on the Hope Fault and Wellington Fault) and suggestions for optimizing future surveys is forthcoming [Langridge et al., in prep].

It is worth mentioning that even obvious fault traces several meters wide, several meters high, and over a hundred meters long can be hard to follow and map in the field (Figure S3.2). Thick understory, downed trees and logs often reduce visibility to ~5 m. Fault traces are often not obvious unless one is directly on them, and even then they can be ambiguous without a careful examination along-strike. At present the positional accuracy of modern handheld GPSs in this environment is ~± 10 m at best. Although structures have been identified in the field below the resolution of the present LiDAR technology and field checking is always recommended, LiDAR has proven itself to be a successful means of imaging the ground surface beneath dense rainforest. The success of this project suggests collection of airborne LiDAR over other portions of the Alpine Fault will likely foster new interpretations of the tectonic geomorphology and highlight the along-strike differences in surface expression.
Figure S3.1  Overview of the 2010 airborne light ranging and detection (LiDAR) survey flown along the Alpine Fault near Franz Josef and Whataroa townships. Northwest-illuminated hillshade (grayscale) derived from 2 m LiDAR digital elevation model (DEM) overlain on hillshade derived from 25 m LINZ DEM data (green shades). The small strip flown into the Whataroa River valley was collected for other purposes.
**Figure S3.2** Four panel comparison of the same area near Franz Josef comparing a 1:50,000 LINZ topographic map with 20 m contours, a hillshade derived from 25 m LINZ DEM data, a high resolution aerial orthophoto and a hillshade derived from the 2 m LiDAR DEM. Notice the geomorphology beneath the thick podocarp rainforest revealed by the LiDAR data. The topographic map and orthophoto were both utilized in identifying cultural features (roads, trails, etc.) present in the LiDAR data to avoid confusion with natural features. Width of view is ~1.4 km. North is towards the top of the page. See Figure 3.4C for Alpine Fault interpretation of this same area.

**S3.2 Characteristics of Fault Traces on the Southern Alpine Fault**

Here, a comparison is made between the characteristics of the fault traces on the southern Alpine Fault and those mapped from LiDAR data on the central Alpine Fault (Chapter 3), to illustrate the differences in fault style as the obliquity of the slip vector varies. Many previous workers have mapped traces of the southern Alpine Fault, mostly in regional reconnaissance studies utilizing aerial photography and limited field examinations [Wellman and Willet, 1942; Clark and Wellman, 1959; Wellman and Wilson, 1964; Hull and Berryman, 1986; Berryman et al., 1992; Sutherland, 1995; Campbell, 2005; Sutherland et al., 2006], but a few more detailed studies exist as well [Cooper and Norris, 1990; Sutherland and Norris, 1995].

During the present study the surface expression of the fault was independently mapped from the Arawhata River to the John O’Groats River using stereographically paired aerial imagery,
Figure S3.3 Two views looking along Alpine Fault traces partially obscured by rainforest vegetation near McKenzie Creek, south of Big Bay. (A) A recent slip (foreground) allows a rare ~25 m view along the Alpine Fault trace. (B) A more typical view along a trace of the Alpine Fault. Other than a particular abundance of tree ferns, slight absence of canopy-forming trees and a slight topographic depression, the fault trace is not obvious. Comparable traces are hard to follow and even harder to map in the field. Airborne LiDAR has proved successful in allowing a bare-earth view of subtle topographic features in this landscape.
oblique aerial photography (e.g., Frontispiece, chapter introductions), and other orthophotography. In addition, the present study is the first to utilize color orthophotography overlain on a three dimensional DEM (e.g., Google Earth), which was found to offer improved resolution of geomorphic features above the existing aerial photograph imagery in several locations. After this first pass of interpretation, comparison was made to the geomorphic interpretations of previous studies, then fieldwork in this study were used to field check and refine the interpretations. The resulting fault scarp interpretations are given as a digital Google Earth .kmz file on the DVD and in Figure S3.4. The following is a brief description of the Alpine Fault scarps from the Arawhata River (north) to the John O’Groats River in the south (refer to Figure S3.4; some detailed place names are shown on 1:50000 LINZ topographic maps).

S3.2.1 Description of Alpine Fault traces from the Arawhata River to the John O’Groats River

Despite a well-defined hillslope break on the northwestern side of the Jackson River, I find field evidence does not support previous interpretations of a fault scarp here [e.g., Hull and Berryman, 1986]. This lineament is likely glacial in origin. Greenland Group rocks exposed at the Arawhata River bridge and along the Jackson River Road near the Arawhata confluence exhibit no hydrothermal alteration and lack dense fractures or faults, implying a distance of >600 m to the Alpine Fault if it is assumed these sort of features are uniformly developed along the entire fault length to the extent observed at the Martyr River. From the Arawhata River to the Martyr River the fault trace is poorly defined, but appears to have a very straight, southeast-side up trace along the southeast side of the Jackson valley where it cuts across nearby the base of steep, incised creeks.

An outcrop on the true right bank of the Martyr River has slickenlines on a fault plane and sediments that are juxtaposed in a way that indicates dextral-reverse motion, consistent with a faint trace in the bush to the northeast. The most obvious bedrock scarp on the northwest side of the right bend in the river does not correlate to a fault trace and occurs within Greenland Group mylonites (presumably the scarp is erosion-related). On the true left of the river the fault trace follows a symmetrically-profiled gully to as far as another outcrop in Fault Creek (the major tributary of the Martyr following the fault). Here (and almost everywhere southwest), the fault exhibits obvious dextral-normal motion. Approaching the saddle into the Cascade River, the flowing creek gives way to poorly drained swampland. The fault has a
very straight fault trace (~059°) from the Arawhata to the Cascade rock avalanche (Chapter 4); along this length the fault appears unsegmented and is characterized by a single trace.

The Cascade River region has the most complex expression of the onshore southern Alpine Fault. The most prominent trace of the fault takes a left-stepping bend southwest of the Cascade rock avalanche and is defined by structures characteristic of strike-slip faults (e.g., pressure ridges, sag ponds, dextrally offset creeks and fan surfaces; Figure 1.0). An outcrop on the true right of Woodhen Creek has slickenlines indicating dextral > normal motion. The hillside opposite (true left) has a complex array of southeast-facing scarps that become harder

Figure S3.4  Alpine Fault traces (in red) in South Westland (from the Arawhata River in the north to the John O’Groats River in the south) mapped in this study (after previous work referenced in the text) overlain on aerial orthophotography (source: Google Earth). The bolder red lines indicate the dominant geomorphologically-defined trace of the Alpine Fault. The surface rupture pattern here is distinct to that on the central Alpine Fault (Figure 3.2B–C).


to follow to the southwest until Woodhen Pond. The most obvious trace of the Alpine Fault bends to a more westerly strike along a well-defined fault-parallel gully, rejoining the Cascade River above a prominent gorge. The fault does not outcrop in the Spoon Slip drainage, which indicates its position is nearer to the Cascade River than this feature. This trace and another prominent northwest-side-up trace to the south through Theta Tarn were recognized by Berryman et al. [1992] to form a fault-bounded lozenge. The present study suggests another trace between these two fault traces. Left-stepping and right-stepping bends in the Cascade are some of the largest indentures in the linear trace of the Alpine Fault; the present study suggests it is possible a fault may follow the Cascade River here as well. This fault is inferred to be the lithologically-defined plate boundary, with the more obvious fault traces to the southeast having reactivated pre-existing boundaries between Pacific plate lithologies.

The two prominent traces defining the fault-bounded lozenge merge again in the Duncan River catchment where the trace is once again straight, with an 052° strike. A prominent 100 m-wide pull-apart basin occurs at the saddle between the Duncan River and Low Creek (e.g., Frontispiece). Well-defined shutter ridges indicating ~200 m dextral offset (with a northwest-side-up throw) occur along the fault near Low Creek. Southwest of the saddle with the Jerry River the fault becomes considerably harder to map. Southwest of the Jerry River is a complex array of ~400 m-wide fault traces with en echelon traces between (also mapped by Bishop, 1995). This is another area where LiDAR data would greatly improve mapping. Two 200 m-spaced traces exist in the lower Durwards Creek area; one of these has likely been enhanced by an abandoned c. 50 yr-old Cascade-Pyke Road.

From the Pyke to the southwestern side of Lake McKerrow, the fault has a remarkably straight (050° strike), single well-defined northwest-side-up fault scarp with several obvious dextral offsets across it (Chapter 5). Single, dry up-to-the-northwest gullies along the fault were observed at McKenzie Creek and Hokuri Creek areas. On the southwest side of Lake McKerrow, unfractured and unaltered St Anne Formation outcrops in the prominent creek and the fault instead follows a smaller parallel tributary. The saddle here is swampy and poorly-drained.

The Alpine Fault does not have a well-defined trace through much of the Kaipo catchment, but can be projected between the offset hill (shutter ridge) near Butchers Creek and hillslope at the Kaipo Slip (e.g., Figure 5.8). South of the Kaipo Slips, the fault is on the northwest side of the prominent fault-parallel creek with unaltered St Anne Formation outcropping in the creek bed. There may be a ~100 m step-over on the saddle between the Kaipo and John
O’Groats rivers. These step-overs seem to be common on saddles in South Westland. A well-defined up-to-the-northwest scarp stretches much of the 2 km-width of the floor of the Kaipo valley. A ~300 m-amplitude left-stepping bend has been mapped by Cooper and Norris [1990] associated with en echelon thrust, normal and symmetrical gully-defined fault traces (and pressure ridges) at the constraining bend in the fault. An up-to-the-northwest uplifted shore platform is present where the fault continues offshore. This trace is presumed to link to the Alpine Fault trace at the mouth of Milford Sound mapped in detailed bathymetry by Barnes et al. [2005].

S3.2.2 Comparison to central Alpine Fault

Although there currently is no LiDAR data on the southern Alpine Fault with which to directly compare to the central Alpine Fault, field checking of southern Alpine Fault trace interpretations in the present study gives a reasonable level of confidence with which to compare the geomorphic expressions. In the central Alpine Fault, it was observed that steeply-dipping dextral-normal faults seemed to produce sharper, more stable (less likely to collapse) fault scarps than shallower dipping (oblique-) thrust-sense scarps. Because the Alpine Fault plane southwest of the Martyr River is generally steeply-dipping (~70–90°; Figure 2.3A) and is associated with a slight normal component of offset, southern Alpine Fault fault traces are generally better-defined than the central Alpine Fault. Often the southern Alpine Fault traces are associated with characteristic strike-slip morphology (e.g., pressure ridges, shutter ridges, sag ponds, strike-slip offsets) rare on the central Alpine Fault.

As discussed in Chapter 3, interaction of a moderately-dipping, non-optimally oriented fault plane accommodating oblique motion at depth with the near-surface processes of fluvial incision, topography and sediment interaction, can cause serial and parallel partitioning to arise in the near-surface. Because of this partitioning, there is almost never a single fault trace at any given location along the central Alpine Fault. Instead, a ~100–1000 m-wide zone of (oblique-) thrust faults and dextral-normal faults develop. Although dextral-normal traces are the best-defined, most movement is inferred to have occurred across the rangefront (oblique-) thrust traces. The orientations of fault traces here also vary considerably due to shallow partitioning (Figure 3.3).

On the steeply-dipping, better oriented (with respect to the modern stress field for slip) southern Alpine Fault, the zone of surface rupturing is considerably more localized, commonly on a single trace. The orientations of fault traces are comparably (to the central
Alpine Fault) much less variable, and in most places strike between 050–060°. The region of
greatest complexity occurs in the Cascade River, where fault-bounded slivers of primarily
metavolcanic and intrusive rocks are present between the Alpine Fault and a comparably
unfaulted (apart from topographically rooted slope failures) mass of Dun Mountain
peridotites and serpentinites. Together these reactivated traces produce a complex
transtensional fault-bounded lozenge structure. Small-scale step-overs on the order of 100–
400 m are inferred to root into a single fault plane at depth. Curiously, steep fault-parallel
drainages commonly cut into unaltered and unfractured bedrock with the Alpine Fault
traversing 30–150 m laterally associated with a less well-defined drainage or gully (instead of
directly in the drainage’s lowest areas). The shallow depth to bedrock in most areas on the
southern Alpine Fault likely also plays a significant role in the minimal width of its surface-
rupturing zone; a wider surface rupture zone might be expected if the steeply-dipping strike-
slip-dominant fault were buried beneath considerable thickness of sediment [cf. Darfield
earthquake surface rupture — Quigley et al., 2011].
S4.1 Analogs for a Large to Great Alpine Fault Earthquake

Likely analogous scenarios to a large or great Alpine Fault earthquake are the 1999 Mw 7.6 Chi-Chi earthquake in Taiwan, the 2002 M 7.9 Denali earthquake in Alaska, and the 2008 Mw 7.9 Wenchuan earthquake in China, which all occurred in tectonically active mountain belts with steep slopes and high rainfalls.

Greater than 20,000 landslides were triggered by the Chi-Chi earthquake, 56% of these landslides were reactivated in a subsequent high rainfall event, and suspended sediment concentrations in nearby rivers were as much as four times higher than background values in the two years following the earthquake [Dadson et al., 2004]. Similar post-seismic landscape effects on the South Island are likely to persist for at least decades after an Alpine Fault earthquake [e.g., Howarth et al., 2012]. The largest failure generated by the Chi-Chi earthquake was the ~0.15 km$^3$ Tsaoling slide which had pre-existing sackung-like depressions at its headscarp prior to failure and has been the site of several historical failures including ones in 1862 (earthquake-triggered), 1941 (earthquake), 1942 (rain-triggered) and 1979 (rain) [Hung, 2000; Chigira et al., 2003]. Slope failures destroyed the Central Cross Island Highway, an important thoroughfare through the mountainous region of the country [Hung, 2000]. In many locations, earthquake shaking resulted in total stripping of the forest cover on many hillslopes [Hung, 2000]. The post-seismic visual impact alone from landslide scars, stripped forests, and of muddy, debris-choked, sediment-laden rivers is likely to have a lasting effect on the tourism industry on the West Coast of the South Island.

The 2002 M7.9 Denali earthquake triggered thousands of landslides, the vast majority of which were confined to a narrow band about 30-km wide that straddled the fault rupture for more than 300 km [Jibson et al., 2004]. The largest rock avalanches clustered near the region of the rupture where ground acceleration and ground-shaking frequencies were likely the highest [Jibson et al., 2004]. The largest rock avalanches (up to 0.02 km$^3$) were deposited onto glaciers, which allowed them to travel distances up to ~11 km during initial deposition [Jibson et al., 2006]. The three Black Rapids rock avalanches were located within 2 kilometers of each other and all had deep-seated bedrock failures extending down much of the hillslope and spread out onto glacier in a wide valley [Jibson et al., 2006]; these failures are most similar in
morphology to the Cascade rock avalanche (CRA) and Round Top (RT) rock avalanche in the Southern Alps. The 2002 Alaskan earthquake triggered rock avalanches with particularly high vertical drops (~0.7–1.65 km) [Jibson et al., 2006], and may be more typical of failures expected around the main divide of the Southern Alps than on its lower elevation western rangefront.

The 2008 Mw 7.9 Wenchuan earthquake was particularly disastrous, triggering more than 56,000 landslides [Dai et al., 2011], which directly caused more than 20,000 fatalities [Yin et al., 2009]. Earthquake-triggered landslides extensively damaged essential lifelines such as highways, bridges, housing and irrigation networks, and many settlements became isolated [Tang et al., 2011]. Coseismic landslides dammed watercourses to create 34 large lakes that threatened many who lived downstream [Cui et al., 2009]. Over 130 large coseismic landslides comprised ~50% (or > 15.5 km²) of the total area of coseismic landslides [Tang et al., 2011], including the Daguangbao landslide with an estimated volume of ~0.84 km³ and a runout of 3.5 km [Chigira et al., 2010]. While the West Coast of the South Island is fortunate to be under-populated with limited infrastructure, these case studies should serve as palpable warnings of both the immediate and lasting effects of fault-proximal landscape response.

S4.2 Geography of the Cascade River Valley

A description of the major geomorphic features of the Cascade River valley is given to provide context to the Cascade rock avalanche (CRA) and other noteworthy slope failures. The Cascade River valley is one of the major catchments crossing the Alpine Fault in South Westland (total catchment area: 440 km²; catchment area upriver of the CRA: 235 km²) (Figure S4.1). The catchment is highly asymmetric with the tributaries entering on the true right (cutting through ultramafic and schist lithologies) contributing to the majority of the catchment area. Many of these tributaries were formerly glaciated hanging valleys. Remnant glaciers still exist on the south-facing heads of many of these valleys. For much of its length from the ~2000 m main divide, the Cascade River is a high-gradient mountain river. In its upper and middle reaches the river runs through several narrow bouldery bedrock gorges and at one point the entire river flows over the 50 m high Durwards Falls.

S4.2.1 Middle Cascade River valley

In its middle reaches, the Cascade valley follows the Alpine Fault for a distance of ~11 km (Figure 4.1B). Southwest of Mt Delta, Korup [2004] identified the very large Hope Blue River
Range (HBRR) rock avalanche formed in Greenland Group metasediments northwest of the Alpine Fault (shown on Figure 4.1B). The raised area of irregular ridges and ponds north of the junction of the Cascade River and McKay Creek have ultramafic detritus, stunted vegetation in places, and subhorizontal ridge crests which argue for a glacial deposit here (Figure 4.1B). An 11 m vertical section of laminated lacustrine silt outcropping in an up-to-the-northwest fault scarp at the junction of the Cascade River and McKay Creek could be a product of ponding behind either the HBRR rock avalanche or a terminal moraine.

The inundation of the entire cross-section of the Cascade valley floor by the HBRR rock avalanche debris suggests the river was dammed by the event. The resulting dam height would have impounded a reservoir extending ~10 km upvalley. The Cascade River cuts a gorge through the rock avalanche deposit and into the underlying in situ Greenland Group, which outcrops on river left. Locally-derived 1–10 m scale angular-subangular Greenland Group boulders line the riverbanks of this gorge and are likely rock avalanche-derived. Similar Greenland Group boulders line the true left bank of Woodhen Creek below Pts 168 and 151; their source is the Hope Blue River Range (outside of the Woodhen Creek catchment) and a rock avalanche origin is suggested. On geomorphologic grounds, the HBRR rock avalanche appears significantly older than the CRA; the HBRR deposit has poorer feature preservation, a more deeply incised river gorge, and a better-defined Alpine Fault scarp crossing it.

The valley from Woodhen Creek to the CRA is dominated by active and inactive ultramafic-derived alluvial fans originating from the Olivine Range on the southeast side of the Cascade valley, and displays some of the best-preserved tectonic geomorphology anywhere along the Alpine Fault. Five elongate hills immediately northwest of the most obvious fault trace (Pts 168, 151, 122, 144 and unnamed hill north-northeast of 144) are likely pressure ridges consistent with a northwest-side-up 1.8 km wide compressional jog in the Alpine Fault here. This represents the largest deflection in the linear trace of the Alpine Fault for over 100 km. A geomorphic and geometric argument is made to infer that the lithologically-defined plate boundary may in fact be along the Cascade River here (Figure 4.1B). Woodhen Pond and two other ponds near Pt 144 have ponded against northwest-side-up fault scarps. Many of the creeks here show obvious dextral deflections, though dextrally-offset features are not obvious. Up-to-the-northwest fault scarps are prominent here. An outcrop of the Alpine Fault cutting an abandoned alluvial fan on the true right of Woodhen Creek has a crudely-
Figure S4.1 Shaded area defines the Cascade River watershed overlain on a northwest-illuminated hillshade derived from 15 m LINZ data. The Alpine Fault forms the prominent lineament striking diagonally to the top right. Notice the high asymmetry of its catchment area in the upper and middle reaches, the effects of glacial deposits, rock avalanches and Alpine Fault offset in its middle reaches, and the way that glacial deposits (particularly on its north side) control its shape in the lower reaches.
developed gouge consisting of glacial till with a single striated principal slip surface indicating a steeply southeast-dipping fault with dextral >> normal motion.

In the Woodhen Creek area, a well-preserved alluvial fan surface opposite Pts 151 and 122 has been abandoned by downcutting of its creek. The front of this fan surface has been truncated by Woodhen Creek to form a 40 m high cliff face of alluvial gravels. Two probable formation mechanisms are proposed: (1) The CRA created a dammed lake that extended upstream of this point, which cut off erosion from Woodhen Creek and allowed the fan to significantly prograde into the lake. Subsequently the Cascade River breached the avalanche dam, cut a gorge, and sought a lower equilibrium base level affecting drainages upstream. (2) The alluvial fan surface dates to a time when Woodhen Creek flowed through its now abandoned channel between Woodhen Pond and Pt 168, allowing alluvial fans along the present lower reaches of Woodhen Creek to propagate unhindered. The lack of preserved alluvial fan surfaces at the next creek to the south and the young appearance of the surface argue for the former mechanism, but more data are needed.

**S4.2.2 Lower Cascade River valley**

From where the Cascade River exits the Cascade rock avalanche deposit at The Bend, the valley widens dramatically to 3–5 km and the river shallows in gradient (total drop of 15 m along 30 km of reach) and displays highly-sinuous kilometer-scale meanders. The lower Cascade valley floodplain is poorly-drained and contains extensive swamplands. The Cascade River is tidally influenced for up to 18 km upstream of its mouth. An abandoned sea cliff extends from Cascade Point 1.5 km inland towards Lake Jumbuck, further evidence of recent progradational sedimentation. Early Holocene sand and silt deposits containing marine bivalves outcrop near the junction of the Cascade and Barn rivers, 3 km from the present coastline [Sutherland, 1995]. Northeast of this location near Pt 12 is a remnant of a longitudinal coastal dune ridge complex, presumably of a comparable age to the marine deposit, considering its proximal location. These dunes are likely comparable to extensive coastal dune complexes studied by Wells and Goff [2006, 2007] that relate to large volume river-transported sediment pulses following major Alpine Fault earthquakes [Wells and Goff, 2007], with large slope failures playing a primary role in this post-seismic sediment response [Korup et al., 2004].

Dextral movement on the Alpine Fault has allowed the Cascade River to capture the Martyr River from the northeast-flowing Jackson River via the youthful Monkey Puzzle Gorge. The
Martyr River was formerly the headwater of the Jackson River until about 1.2 km of dextral offset ago (this would have been c. 44,000 yrs ago at 27 mm/yr strike-slip rate). Since no other significant tributaries are found on the Pacific plate in the 7.5 km to the southwest, it seems likely the Monkey Puzzle Gorge has been carved by the Martyr River within the last 50,000 yrs. Vertical 50 m high exposures of iron-cemented fluvial-glacial gravels line portions of Monkey Puzzle Gorge indicating the gorge has been in-filled and subsequently re-carved at least once. A similar but smaller gorge 1 km north of the Martyr River cuts 80 m through a glacial bench at 100 m elevation that was covered in ice during the last glacial maximum (LGM). With the exception of the Martyr capture, there have been no other major changes in the Cascade catchment in the last c. 300,000 yrs.

The Cascade Plateau, a vast area of lateral moraines at the mouth of the Cascade River, preserves a record of glaciations extending back over 117,000 yrs [Sutherland et al., 2007b]. Many of the glacial trim lines and lateral moraine crests can be traced from the Cascade Plateau to the Martyr River and indicate that glacial deposits between the Cascade and Martyr rivers are late LGM (≤ 19 ka) in age [Sutherland, 1995; Rattenbury et al., 2010]. The upstream terminus of these moraine crests is immediately northeast of the Cascade rock avalanche. Charlies Bump is a 70 m high bedrock knoll in the center of the Cascade valley that has been glacially scoured; similar bedrock hills protrude through moraine northeast of the Cascade rock avalanche. As well as providing an upper limit on the age of the Cascade rock avalanche, LGM lateral deposits reach an elevation of > 500 m at the Cascade Plateau, which indicates the LGM Cascade glacier must have reached a similar elevation in the vicinity of the Cascade rock avalanche and most, if not all, of the valley wall would have been glacially scoured at this time. Valley walls have been glacially oversteepened and slopes in excess of 30° are not uncommon.

S4.3 Lithology and Sackungen

Sackungen are ubiquitous along Martyr Spur in Dun Mountain Ophiolite Group serpentinites, but are comparatively uncommon in the less serpentinized harzburgite south of Woodhen Creek. The Dun Mountain Ultramafic Group serpentinites do not have a pervasive planar fabric like the nearby schists, but they do contain steeply-dipping anastomosing serpentine shear zones cross-cutting the serpentinites at intervals. These serpentine shear zones are likely to provide weaknesses in the rock mass of Martyr Spur, but the sackungen traces do not follow these serpentine shear zones. Together these observations suggest the presence of weaker rock may encourage sackungen formation, but that the rock anisotropy
does not significantly influence sackungen orientation here (the latter point in agreement with Beck’s 1968 observations for schist and greywacke elsewhere in the Southern Alps).

Elsewhere in the study area, sackung has been identified cutting all lithologies east of the Alpine Fault. It is presumed that the lower elevation and lack of glacial oversteepening of the ranges west of the Alpine Fault explains the lack of sackung there. Sackung is particularly abundant on the northern Red Hills Range between the Duncan and Cascade rivers (Figure S4.2), where they generally indicate asymmetric collapse towards the Cascade. Sackung is also prominent on the Skippers Range, where they are often difficult to distinguish from active faults, which also cross the range.

![Figure S4.2](image)

Well-defined uphill-facing sackung on the northern Red Hills Range between the Duncan and Cascade rivers indicating asymmetric collapse towards the Cascade (to the left of this photo). View is looking southwest. Scarp heights are typically 1–5 m. Notice the en echelon character of the scarps and the prominent scarp below bushline (center right).

**S4.4 The Pyke Rock Avalanche**

In the course of fieldwork a 80 m-long outcrop of a rock avalanche deposit was discovered in the right bank of Durwards Creek, about 800 m upstream of its junction with the Pyke River.
The outcrop exhibits characteristic rock avalanche textures: highly-fractured jigsaw-textured metavolcaniclastic rock forming large meter-scale semi-coherent blocks of formerly intact rock (Figure S4.3); significant dilation of the rock mass is indicated. A gently undulose contact caps a 3–4 m section of the rock avalanche deposit with overlying fluvial gravels or glacial till. Because of this, the landscape immediately surrounding the outcrop is flat and there is no obvious surface expression of the rock avalanche deposit.

![Figure S4.3](image)

**Figure S4.3** Outcrop of the Pyke rock avalanche deposit in Durwards Creek (see Figure S4.4A for location). Height of the outcrop is ~8 m. Here a 4–7 m section of rock avalanche deposit is overlain by 1–4 m of glacial till (top left), which at least partially includes reworked rock avalanche deposit. View looking northwest.

By examining aerial photographs I have examination identified a headscarp defined by the headwaters of Chrome Creek and the unnamed creek to the southeast. A large mass of rock below the scarp bulges into the junction of the Pyke and Durwards valleys; this failure is herein named the Pyke rock avalanche (Figure S4.4). During previous 1:250,000 scale geologic mapping the area was delineated as Quaternary glacial till [Turnbull, 2000]. The top of the headscarp is at ~900 m elevation while the outcrop examined is at ~80 m indicating a minimum vertical fall ($H_{\text{max}}$) of 820 m. The horizontal distance between these two points
indicates a minimum runout distance ($L_{\text{max}}$) of 2600 m. The horizontal distance from the base of the source area to the Alpine Fault is ~1.5 km. The rock avalanche deposit is estimated to have an area of ~2 km$^2$, although erosion may have removed significant portions of the
deposit along the Pyke River and the distal extent of the deposit is unknown. To the author’s knowledge this is the only rock avalanche identified in the region that may have pre-dated the LGM. As emphasized in Chapter 4, rock avalanche deposits are commonly misinterpreted as glacial deposits. Older deposits, in particular, are more likely to have their morphologies or extent altered by erosion and a thin cap of glacial till could obscure an underlying rock avalanche deposit. Although it may be easy to detect a rock avalanche deposit or source area when one knows where to look, during the present study I would not have identified the headscarp and main deposit area if field examination of the deposit outcrop had not stimulated subsequent examination of aerial imagery. The source area and main deposit is not obvious from a topographic map and it was only by an oblique aerial view (e.g., Figure S4.4B) that the source area and main deposit could be confidently identified.

S4.5 Possible Effects of Alpine Fault Permeability

Dynamic strain during an earthquake can locally increase fluid pressure along the fault, potentially enough to be a contributing mechanism for collapse. The Alpine Fault core exposed 1 km northeast of the CRA consists of a 1.5 m thick tabular zone of essentially impermeable (10^{-20} – 10^{-19} m^2) phyllosilicate-rich fault gouge and is likely to act as a barrier to groundwater flow, whereas the rest of the fault-damaged rocks in the source area should exhibit comparably high permeabilities (see Chapter 2). Stable isotopes of vein-hosted carbonate from both sides of the fault core are consistent with the fault core remaining a barrier to hydrologic flow through the seismic cycle, including during earthquake rupture (see Supplement 2). Measurements in the Deep Fault Drilling Project DFDP-1 boreholes on the central Alpine Fault at Gaunt Creek have detected a 0.53 MPa step in fluid pressure measured across the principal slip zone in the shallow ~150 m borehole, confirming the fault core there acts as a hydraulic seal in the present interseismic period [Sutherland et al., 2012]. The permeability of the gouge at Gaunt Creek (k = 10^{-20} m^2) is comparable to that at the Martyr River, suggesting comparable fluid pressure compartmentalization may be realized along considerable strike length of the fault. Cox et al. [2012b] showed that during the Mw 7.1 Darfield (Canterbury) earthquake of 2 September 2010 the fluid pressure in groundwater aquifers increased coseismically in the vicinity of the fault. Taken together, these data suggest the low permeability fault core of the Alpine Fault may have elevated background fluid pressures that can greatly increase coseismically and act as a weakening mechanism in conjunction with ground accelerations. Although speculative, this mechanism is worth exploring.
S4.6 A Hypothetical Temporal Model Linking Glacial Cycles and Large Slope Collapses

The modern Southern Alps / Fiordland are an inherited landscape largely carved by past glaciations [e.g., Barrell, 2011; Shuster et al., 2011] in conjunction with uplift. During interglacial times, fluvial and mass wasting processes attempt to reach a new equilibrium by reshaping the inherited glacial landscape. Sackungen and RSFs are presently widespread through the Southern Alps and indicative of post-glacial topographic adjustment [e.g., Korup, 2005b; Beck, 1968]. Much of the steep, formerly-glaciated western rangefront of the Southern Alps is heavily-dissected by steep, rapidly eroding creeks with active slip heads, such as the hillslope immediately southwest of the CRA (Figure S4.5). These creeks can help prevent large deep-seated slope failure of an oversteepened hillslope by lowering the mean gravitational potential surface of the remaining rock mass, dividing the mass into smaller components, and encouraging smaller, shallower slope failures. If this effect is significant, large deep-seated slope failures should decrease in abundance with time following a glaciation. Where rock avalanches are deposited into narrow valleys, the rivers may be able to remove the deposits over timescales of years to centuries. In wider valleys such as the Cascade, portions of the deposit may be able to lag for thousands to tens of thousands of years. Glaciers are suspected to be more effective at removing glacial deposits than rivers in wide valleys as the glacier stretches from wall to wall and can override and excavate any deposit in its path. These valley-spanning glaciers will also erode the ridges between the interglacially-formed steeply eroding creeks (which probably shut down during glaciations) to once again smooth the valley walls. This glaciation widens the valley and once again steepens hillslopes, which then become unstable when the buttressing ice retreats. The mass of ice filling a valley buttresses the valley hillslopes, and from a gravitational point of view, effectively raises the elevation of the valley floor. Thus, large deep-seated rock avalanches rooted into the valley floor in the middle and lower reaches of a valley (like the CRA or the Green Lake landslide) are expected to be less prevalent during glaciations. By this reasoning, during glaciations catastrophic failures may preferentially focus at the heads of steeply glaciated valleys where they fall off steep slopes, but do not necessarily root into the base of the topography. Comparable examples would be the modern rock avalanches and rock falls near Mt Cook. While earthquake triggers occur irrespective of a particular time in a glacial cycle, it is argued that conditions for very large deep-seated catastrophic failures like the CRA, RT, HBBR, and John O’Groats are most ideal immediately following glacial retreat.
Figure S4.5  Annotated aerial photograph and slope profiles (as indicated) of the region immediately southwest of the Cascade rock avalanche (CRA) illustrating the effects of post-glacial incision in lowering the average gravitational potential surface of formerly glaciated hillslopes, which conceptually should discourage large scale deep-seated failures with time.
S5.1 Physiography of the Westland Coastal Plains & Hills

For nearly 400 km along the Alpine Fault from near the Ahaura River in the north to where
the Alpine Fault goes offshore near Milford Sound in the south, the land to the west of the
Alpine Fault has a distinct geomorphology unique to New Zealand, and perhaps the world.
This coastal strip (referred to as the Westland Coastal Plains & Hills geomorphic province in
this study) can be as narrow as 6 km (Bruce Bay) or as wide as 25 km (Hokitika) (Figure 5.1B;
see also Figure 1.5). The four main features of this landscape are (1) well-developed alluvial
fans which extend from where major rivers exit the Southern Alps to the Tasman Sea, (2)
progradational coastal dune ridge complexes common from Hokitika to the Hollyford but
particularly well-developed near Haast, (3) extensive Pleistocene glacial moraines and outwash
gravel deposits to 300–400 m elevation, which bound alluvial fans and disrupt drainages to
create an abundance of lakes and lowland swamps, and (4) bedrock “islands” of isolated hills
or discontinuous ranges up to 1350 m in elevation, but typically ~600–900 m. The following
sections provide further detail of these principle features.

S5.1.1 Alluvial fans and coastal dunes

Persistent cycles of aggradation and trenching of the heads of the alluvial fans (i.e., at the
Alpine Fault-controlled western rangefront of the Southern Alps) have been suggested to
result from infrequent catastrophic sediment inputs [Davies and Korup, 2007]. When such a
sediment pulse, commonly attributed to major Alpine Fault earthquakes [e.g., Whitehouse and
Griffiths, 1983; Wells and Goff, 2006, 2007; Howarth et al., 2012] reaches the sea, it can form a
new progradational dune ridge, effectively extending the width of the coastal plain. LiDAR
imagery shows no evidence for a 1717 AD Alpine Fault earthquake scarp (or any previous
events) crossing any of these wide alluvial fans, indicating the entire surface is active on
centurial to millennial timescales (Chapter 3). The current long-term aggradation rate for the
Waiho fan is ~2 mm/yr [Barrell et al., 2003]; aggradation rates for other major fans are likely
comparable.
S5.1.2 Glacial deposits

As noted by Suggate [2004], the extent to which features related to glacial records from the last c. 500 kyr are preserved in New Zealand varies widely depending on the tectonic situation and the ease with which younger ice advances and fluvial erosion can obliterate evidence of older ones. In North Westland, the area between the Hokitika and Grey Rivers is considered the type area for Middle to Late Pleistocene glaciations in New Zealand due to well-preserved morpho-stratigraphic relations between glacial outwash surfaces and uplifted beach terraces which allow correlations of glaciations back to marine isotope stage (MIS) 8 at c. 250 ka [Suggate, 1965, 1990, 2004; Suggate and Waight, 1999]. Drill cores from offshore the east coast of the South Island record nine glaciations in the last 700 ka, but only evidence for the last four glaciations (back to c. 350 ka) is preserved onshore due to uplift and erosion [Suggate, 1990]. Where these glaciers crossed the Alpine Fault and exited the narrow confines of the Southern Alps valleys onto the coastal plain in central Westland, they spread out to form distributary-lobed piedmont glaciers, as evidenced by preserved moraines. Characteristically, many of the major glaciers formed a pattern of three arcuate, westerly-bending lobes (e.g., Waiho, Whataroa, Wanganui) that may form partly as a function of interfering bedrock hills and high rates of dextral displacement on the Alpine Fault. Remnants of older ice advance landforms (e.g., between 150–350 ka) deposited on the Australian (AUS) plate tend to be best preserved on the northern side of a glacial system, due to dextral motion on the Alpine Fault at the rangefront [Barrell, 2011]. In the north the continental shelf is wide and all last glacial maximum (LGM) glaciers probably terminated onland during glacio-eustatic sea-level lowstands, whereas to the south some glaciers may have had tidewater calving termini (e.g., Paringa; Barrell, 2011). Many moraines and outwash surfaces in this province terminate in cliffs at the coast, indicating their seaward extent was formerly greater.

While glaciers rooted west of the Southern Alps are rare here owing in part to the general low elevation and limited catchment development on the isolated hills, glacial cirques and U-shaped valleys are prominent features of the Hohonu Range east of Hokitika and on southeastern sides of nearby hills of similar elevation [Barrell, 2011]. Similarly, on the AUS plate south of Jackson Bay there is evidence for small glaciers on the southeastern slopes of the McArthur Tops, Hope Blue River Range, and McKenzie Range, and potentially within the west-flowing Hope River catchment.
In general, glacial deposits south of Jackson Bay have received very little attention compared to further north. Regional reconnaissance work on glacial deposits was carried out by Turner [1930b] and Sutherland [1995]. Detailed work in this region has focused on the Cascade Plateau lateral moraine sequence north of the mouth of the Cascade River where $^{10}$Be cosmogenic nuclide ages between 117–14 ka have been recorded [Sutherland et al., 1995, 2007b], and at Hokuri Creek where the ages of Alpine Fault offset moraines were correlated against the North Westland sequence and used to determine strike-slip rates [Sutherland and Norris, 1995; Sutherland et al., 2006].

**S5.1.3 Bedrock islands**

Basement rocks west of the Alpine Fault are collectively referred to as the Western Province; this comprises Ordovician Greenland Group metasandstones and metamudstones of the Buller terrane and intruding mid-Paleozoic granitic rocks of the Paringa and Karamea suites which together make up the isolated hills and round ranges characteristic of this part of Westland [Rattenbury et al., 2010]. Along the southern 170 km of AUS plate coastline, Tertiary limestones, sandstones, mudstones and some basalt crop out as a northwest-dipping limb of a coastal monocline [Rattenbury et al., 2010].

Isolated “beehive-shaped hills” (e.g., Mt McLean, Mosquito Hill, Ralphs Knob) are the most conspicuous geomorphic features of the coastal plain where they protrude from veneers of Quaternary alluvial/fluvial and glacial sediments to elevations in excess of ~700 m, often at significant distance from the rangefront. The rounded summits of these hills indicate they were likely overrun by glaciers and thus attest to the depth and extent of former glaciers beyond the Southern Alps. Wellman and Willett [1942] postulated that these hills were composed of more resistant granitic rocks intruding a weaker metasedimentary sequence and formed through a combination of differential erosion of the varying bedrocks and glacial erosion. It has since been shown many of these hills in fact are composed of Paleozoic Greenland Group metasedimentary rocks and that there is no obvious differential erosion between these rocks and the plutonics [e.g., Ryland, 2008; Rattenbury et al., 2010]. The present study advances a conceptual model to explain how these features result from a combination of dextral offset on the Alpine Fault and multiple glaciations.
S5.1.4 Australian plate deformation

The Westland Coastal Plains & Hills province is relatively unique in New Zealand in that it is relatively devoid of faults and, with the exception of one small fault at Smithy Creek 5 km west of Franz Josef [Lund-Snee, 2011] and the Bald Hill Thrust south of Hokitika [Rattenbury, 1986], has no known active faults onshore (e.g., Figure 1.4). Widespread glacial moraines and glaciated hillslopes form coast-perpendicular strain markers that nowhere record post-glacial offset other than at the Alpine Fault; a particularly good example is at the Cascade River where lateral moraines and glacial trim lines can be traced over 20 km from the coast to the Alpine Fault with no fault disruption. Similarly, current seismicity [e.g., Eberhart-Phillips, 1995; GeoNet, 2012] and geodetic strain [Beavan et al., 1999] are both concentrated east of the Alpine Fault here. Other than the South Westland coastal monocline, evidence for Quaternary folding is often localized very near the Alpine Fault [e.g., Simpson et al., 1994; Sutherland and Norris, 1995]. In short, there appears to be very little Post-Pliocene internal deformation of the onshore AUS plate south of the Ahaura River.

Flights of uplifted marine terrace surfaces are widespread in central Westland both on the AUS plate at elevations to ~300 m and as remnants notching the rangefront of the Southern Alps up to heights of ~1500 m. Correlation with dated terraces in New Guinea [e.g., Bloom et al., 1974] yield inferred ages up to c. 350 ka for the highest terraces preserved on the PAC plate; this correlation gives localized PAC plate uplift rates of ~3–6 mm/yr near Kaniere in the north, ~3–8 mm/yr near Franz Josef, and ~4–6 mm/yr near Haast further south [Bull and Cooper, 1986]. AUS plate vertical motion rates in this region are comparably lower (~0.86–2.0 mm/yr) indicating a net uplift of the PAC plate [Bull and Cooper, 1986; Cooper and Kostro, 2006]. Zircon and apatite fission track studies on Greenland Group rocks between Jackson Bay and the Taramakau River by Seward and Nathan [1990] and Kamp et al. [1992] provide useful longer-term AUS plate uplift rates over the last 5 Ma which yield rates of ~2 mm/yr near the fault and < 1 mm/yr along the coast, in agreement with raised Quaternary marine and river terraces [e.g., Bull and Cooper, 1986; Suggate, 1992; Cooper and Kostro, 2006].

Eighteen PAC plate uplifted terraces at Franz Josef span c. 40–330 ka, indicating that during much of this time the western rangefront of the Southern Alps would have plunged more-or-less directly into the Tasman Sea and the major valleys would have been fjords [Bull and Cooper, 1986]. Marine deposits dated at c. 16 ka immediately east of the Alpine Fault at the
Paringa River support deposition in a fjord setting and give uplift rates of \( \sim 7-8 \text{ mm/yr} \) [Simpson et al., 1994].

In summary, the Westland Coastal Plains & Hills province is not deforming internally and is everywhere being uplifted, but appears to have been experiencing higher uplift rates closer to the Alpine Fault and lower rates toward the sea. It essentially acts as a rigid indenter with respect to the thermally weakened PAC plate which has a wide zone of deformation extending to the east coast in places.

**S5.2 The Australian Plate in South Westland**

In contrast to the central Alpine Fault, tectonic geomorphology of the southern Alpine Fault suggests Late Quaternary net uplift on the AUS plate, despite the overall lower elevation of ranges west of the fault here compared to those east of the fault (Chapter 2). This uplift pattern is corroborated by the finding of uplifted c. 15.6 ka marine shells at Lake McKerrow, which yield PAC plate uplift of 1.6 ± 0.3 mm/yr and AUS plate uplift rates of 2.2 ± 0.2 mm/yr [Norris and Cooper, 2001]. The edge of the continental shelf between Jackson Bay and Milford Sound, which is on average 10 km northwest of the present coastline, delineates the approximate position of the coastline during glacio-eustatic sea level low-stands. The AUS plate continental shelf gives way to oceanic crust near Poison Bay, just southwest of Milford Sound (Figure 5.1).

There are several lines of evidence that indicate the AUS plate here is uplifting at a steady rate along a shallowly southwest-plunging Alpine Fault slip vector and that this block has been and continues to be progressively uplifted from the sea (see Chapter 2). This would indicate the region around Jackson Bay has been sub-aerially exposed the longest while the region near John O’Groats River in the south has most recently emerged from the sea and should display the most youthful characteristics. The elevation of the highest peaks in the AUS plate ranges increases consistently northeastward to the Hope Blue River Range southwest of the Cascade River with the ranges northeast of the Cascade River (past the reversal in uplift polarity; Chapter 2) forming slightly lower elevations. In contrast to the central Alpine Fault, tectonic geomorphology on the southern Alpine Fault suggests Late Quaternary net uplift of the AUS plate (Chapter 2; Supplement 3).
S5.2.1 Watersheds

Watersheds in the area are of essentially 5 types:

(1) Headward propagating, slip-headed, bedrock creeks (e.g., Kaipo Slips, Wolf River, Jerusalem Creek)

(2) Former slip-headed, bedrock creeks that have propagated enough to become branching bedrock rivers (e.g., Ryans Creek, Stafford River), potentially with minor glaciations (e.g., Hope River)

(3) Glaciated major river valleys rooted into ranges east of the Alpine Fault (e.g., Cascade River, Hollyford River, Kaipo River)

(4) Creeks incising into glacial surfaces (e.g., Professor Creek, N. branch of Hokuri Creek, Whiskey Creek, Low Creek)

(5) Drainages presently disrupted (and generally dextrally offset) by the Alpine Fault and/or glacial deposits (e.g., Smoothwater River, Ellery Creek, Dry Awarua River, Pyke River)

Major creeks and rivers in the bedrock ranges west of the Alpine Fault all drain to the northwest consistent with (1) northwest-directed tilting of the ranges due to fault-perpendicular differential uplift, and (2) the northwest side of these ranges being the only ice-free margins of the ranges through glacial cycles. The creeks and rivers with headwaters west of the Alpine Fault here have increasingly youthful characteristics southward. Steep creeks near the Wolf Tablelands in the south are still eroding headwardly into an old, poorly-preserved glacial moraine surface, while complex branching river systems are not seen until the latitude of the Hacket and Ryans rivers. Hill slope angles within these bedrock ranges increase steadily northward, which may be a function of landscape age and size and height of the ranges, but cannot be attributed to lithology (Figure S5.1).

S5.2.2 South Westland peak accordance

In contrast to the ranges of the Southern Alps east of the Alpine Fault which have well-defined peaks, those to the west are connected by ridges of similar elevation. When viewed in profile, the ranges west of the fault form a rather smooth ~5° northwest-dipping skyline (see Frontispiece). This (herein named) South Westland peak accordance is suspected to be a result of erosion of some pre-existing sub-planar surface (be it a marine planation surface or
glacial scour surface or erosion parallel to an existing Greenland Group/Tertiary unconformity) and subsequent northwest tilting.

The ~5° northwest tilt of the ranges and peak accordance is consistent with fission track results on Greenland Group metasediments by Kamp et al. [1992] which show 3.6–3.8 km of uplift near Jackson Head increasing eastward to ~5 km of uplift nearest the fault at the Arawhata River bridge, all within the last 5 Ma. This means the basement/Tertiary contact exposed near the coast should be ~1.3 km above the ground surface near the fault. The highest peak in the area is about 1000 m indicating there has been only ~300 m of erosion into the basement Greenland Group and is consistent with the northwest-tilted peak accordance observed. Offshore the Tertiary/basement contact is flat at ~-4000 m elevation [Rattenbury et al., 2010]. If this depth has been relatively stationary over the last 5 Ma, then it would indicate the South Westland monocline along the coast is less than 5 Ma and the onshore AUS plate deformed as a relatively intact block, apart from a 6° tilt to the northwest reflecting fault perpendicular differential uplift. The lack of widespread glacial cirques west of the fault indicates this area was never extensively glaciated, which partly explains why the ridgelines are not more thoroughly excavated and this peak accordance is preserved. Erosion rates in these ranges appear to be much lower than in the adjacent Southern Alps. If as suspected, the South Westland peak accordance indicates there has been limited erosion of the top of the ranges, and uplift rates have been steady over this time, it indicates the AUS plate presently exposed became emergent from the sea c. 480 ka, given an uplift rate of ~2.5 mm/yr.

### S5.3 Amino Acid Racemization Dating of Uplifted Marine Sediments

Realistic uplift rates of marine sediments at ~400–600 m elevation indicate the shell and wood samples collected are unfortunately too old to be radiocarbon dated. For this reason shells were analyzed utilizing the amino acid racemization dating technique to provide some age control on the uplifted sediments. Amino acid racemization (AAR) is a relative dating technique which can be used to date biological materials (e.g., shell, wood, bone). The AAR dating technique works on the principle that all living organisms maintain their amino acids in an L (left-handed) molecular configuration. When the organism dies its amino acids convert to a D (right-handed) configuration at a rate determined by temperature, humidity and acidity, until equilibrium is reached with an equal number of D and L amino acids. Thus D/L ratios between 0 and 1 can be used to provide a relative age compared to other samples with similar diagenetic histories [e.g., Rutter and Blackwell, 1995]. AAR rates vary by temperature
Figure S5.1: Slope map of South Westland, New Zealand. The map shows the slope angle (in degrees) calculated as the first derivative of the 15 m Digital Elevation Model (DEM) data. The map highlights the Alpine Fault (solid line) and the Pembroke Fault (dashed line). The more oblique dashed line delineates the Glade-Darrans Fault. The map also shows the generally low slope angle (< 10°) of glacial deposits and the smoothly sloping slopes of the bedrock ranges west of the Alpine Fault. The southwestern ranges generally have the lowest slope angles, while the southeastern ranges have the highest slope angles due to their recent uplift from the Tasman Sea.
such that the maximum dateable age range for a site with a mean annual temperature of 30°C is c. 200 ka, extending to c. 2 Ma for sites with a mean annual temperature of 10°C [D. Kaufman pers. comm., 2012].

Where possible *Talochlamys gemmulata* (a species of scallop) were chosen for the analyses as they tended to be common in the sediments, diagenetically stable (do not become chalky or lose their luster as easily as other species), and are fragile enough to be unlikely to be reworked (thus yielding a reliable deposition age). The depositional history of the marine units is interpreted, based on the sedimentary succession, to have been relatively simple: (1) a brief period of deposition in 0–50 m water depth was followed by (2) sub-aerial burial beneath ~50 m of glacial till during which the underlying unit gradually increased in elevation but not burial depth. The samples were completely buried laterally until the last c. 100 ± 100 yrs.

### S5.3.1 Analytical methods

The extent of racemization of amino acids, a measure of fossil age based on the increasing ratio of D to L-amino acids, was determined in fossil molluscs from the New Zealand Quaternary successions by Reverse Phase, High Performance Liquid Chromatography at the University of Wollongong following the methods of *Kaufman and Manley* [1998]. Analyses were undertaken on the total hydrolysable amino acids after hydrolysis for 22 h at 110°C in 8 mol HCl. The analytical procedure involved the pre-column derivatization of DL-amino acids with o-phthalaldehyde (OPA) together with the chiral thiol, N-isobutyryl-L-cysteine (IBLC) to yield fluorescent diastereomeric derivatives of the chiral primary amino acids. Amino acid D/L ratio determinations were undertaken using an Agilent 1100 HPLC with a C-18 column and auto-injector. Results are reported for the amino acids glutamic acid and valine based on peak area calculations (Table 5.1; Appendix).

### S5.3.2 Results

Considered individually, the D/L values can vary considerably at a given site, making them hard to interpret in detail (Appendix). For example, some stratigraphically higher samples at *Madagascar Creek* yield higher D/L values than stratigraphically lower samples despite convincing stratigraphic evidence that the unit is upright and the shells chosen are unlikely to be reworked. However, if analyses for individual formations are pooled, and the associated uncertainties bracketed, the results yield a succession of D/L values consistent with elevation
Figure S5.2  Eustatic sea level curves (top) from Siddall et al. [2006] and deuterium concentrations from the Antarctic EPICA Dome C ice core and calculated temperature curves (bottom) from Jouzel et al. [2007] utilized in this study. Large numbers within lower plot indicate position of marine isotope stages (MIS). Table at bottom gives correlations of MIS stage, New Zealand glaciation/interglaciation names and approximate ages from Barrell [2011].
and paleontological estimates (Table 5.1). In the future I will attempt to calculate an absolute age estimate from these D/L values by modeling temperatures throughout the diagenetic history of the samples. For now, a preliminary age of 191 ± 34 ka has been offered for the Sara Formation based on correlation of amino acid racemization analyses to those in well-dated sequences from the Wanganui Basin (North Island) and eastern Australia [C. Murray-Wallace pers. comm., 2011]. As discussed in Chapter 5, the Wolf Formation presently has the most tightly constrained age of c. 290–270 ka. The Teer Formation appears to be older, as suggested by Sutherland et al. [1995]. Uplift rates were determined considering the sea level curves summarized in Siddall et al. [2006], and glacial events correlated against EPICA Dome C ice core 800 kyr deuterium data from Jouzel et al. [2007] (Figure S5.2).

S5.4 Evidence of Glaciations

Glacial deposits are widespread on the AUS plate in South Westland, especially as extensive lateral moraines on the north side of major valleys and as eroded remnants of previously more extensive moraine and outwash surfaces up to elevations of 900 m (Figure 5.1C). Many of the higher elevation features are poorly-preserved, reflecting their greater age (see Figure S5.2 for MIS stage, name and age correlations of New Zealand glaciations and interglaciations). Although direct dating in this region is currently limited to the youngest and best preserved lateral moraines of the Cascade Plateau (10Be exposure ages spanning c. 117–14 ka) [Sutherland et al., 2007b], it is clear from the elevation of features, degree of preservation, cross-cutting relationships, moraine clast lithologies and the relationship to geomorphic expression of Alpine Fault offset that a partial record of several glaciations exists here (Figure 5.2B). Because the strike-slip rate of the Alpine Fault has been relatively constant for the last > 300 kyr, this rate can be used to assign glacial deposits to specific glaciations (e.g., Table 5.2). Because dextral offset and net uplift of the AUS plate valleys from their PAC plate sources protect glacial deposits from subsequent erosion by rivers and glaciers, this area offers a unique opportunity to study glaciations not as well-preserved elsewhere along the western rangefront of the Southern Alps.

S5.4.1 Late Otira (MIS 2)

Evidence for the most recent Late Otira glaciation is widespread in this portion of South Westland [e.g., Sutherland and Norris, 1995; Sutherland et al., 2006, 2007b; Figure 75.2 of Barrell, 2011]. All major valleys with headwaters east of the Alpine Fault had large valley-spanning glaciers which crossed the Alpine Fault. Only small glaciers were associated with headwaters
west of the Alpine Fault; these were confined to highest elevation, south-facing drainages in the McKenzie Range, Hope Blue River Range and the McArthur Tops. The Hope valley was likely the only major valley with headwaters west of the Alpine Fault that was glaciated; there is no evidence for glaciations in the Hacket River and Ryans Creek [Sutherland, 1995].

Lateral moraines at Cascade Plateau on the north side of the mouth of the Cascade valley have been dated at 19–23 ka by \(^{10}\)Be cosmogenic nuclide methods [Sutherland et al., 2007b]. These moraines end in 100 m vertical cliffs into the sea, indicating the extent of glaciation was beyond the present coastline, perhaps as far as the edge of the continental shelf, a further ~10 km to the west. Recessional post-LGM terminal moraines in the Cascade valley define a distributary-lobed piedmont glacier. Glacial deposits in the Smoothwater/Ellery valley have a provenance indicating a source from the Jackson River (e.g., schist, peridotite, nephrite) and based on their preservation are interpreted as LGM. This indicates Lake Ellery and the glacial deposit-imposed drainage divide in the Smoothwater/Ellery valley is post-LGM in age. Lateral moraines (c. 17 ka) were interpreted by Sutherland and Norris [1995] at Hokuri Creek. Well-preserved glaciated valley hillslopes with trimlines and a recessional terminal moraine in the lower Kaipo valley indicate its glacier extended to a position near the present coastline. Similarly, glaciated hillslopes in the lower John O’Groats valley indicate its glacier extended to a position near the present coastline. Lateral moraines of Yates Point (at ~≤ 150 m elevation) may be LGM or associated with an older glaciation (see Figure 5.1C).

The presence of Late Otira deposits in the Gorge River, Big Bay and Lower Hollyford is particularly ambiguous. Low elevation glacial channels and associated terminal moraines spanning the Gorge and Duncan rivers are likely Late Otira in age based on their preservation and the limited degree of fluvial incision that has cut through these glacial surfaces but potentially may be Early Otira (Figure 5.3). Well-preserved lateral moraines on the north side of Big Bay may be related to an LGM Pyke/Hollyford glacier as suggested by Barrell [2011], but this does not explain why the Early Otira offset glacial hillslope at Pyke Corner has been preserved. Curiously, the Lower Hollyford valley lacks extensive glacial deposits like those still present at the mouths of other major valleys of the region. Evidence of glaciation in the Lower Hollyford is confined to a small flat terrace between the mouth of the Hollyford River and Jerusalem Creek, although this could also be interpreted as an uplifted marine terrace. A possibility which should be considered is that the forced split of the Pyke/Hollyford Glacier at the junction of the Pyke and Lower Hollyford valleys caused the glacier to extend a shorter
distance down both of the valleys than its neighbors and that Martins Bay and/or Big Bay were ice-free. Regardless, a lack of morainal deposits upvalley remains puzzling.

**S5.4.2 Early Otira (MIS 4)**

Evidence for the Early Otira Glaciation is comparatively sparse, largely owing to the strength of the subsequent Late Otira Glaciation [e.g., Suggate, 1990; Barrell, 2011], and the comparatively short time span between them (and thus comparatively little offset of the deposits to protect them from the Late Otira glaciers), yet in central and north Westland Early Otira glacial deposits are some of the largest and most extensive [e.g., Suggate, 1990]. Sutherland et al. [2007b] determined that lateral moraines at the Cascade Plateau date from 80–58 ka. Offset hillslopes and moraines in the ~300 m elevation saddle between the Hollyford valley and Big Bay are the most obvious indications of Early Otira deposits [Sutherland and Norris, 1995; Sutherland et al., 2006; this study].

**S5.4.3 Waimea (MIS 6)**

The most convincing evidence for the Waimea Glaciation in South Westland is the paired lateral moraines of Low Creek which have been dextrally offset by ~2900 m from the Jerry River headwaters (although these may instead correlate to an early Early Otira advance; Figure 5.2B; Figure 5.10). The unique extent to which they are preserved is attributed to the fact that these moraines have not been aligned with a glaciated catchment since they were offset from the Jerry River. From floor to crest, these paired moraines rise over 200 m; considering the small catchment size of the Jerry River, it is suggested this glaciation would have formed deep and extensive piedmont glaciers in the neighboring main valleys. However, the lack of obvious Waimea-aged glacial deposits elsewhere in South Westland suggests the subsequent two glaciations were even more widespread and obliterated most evidence of this glaciation. The flat glacial surface at ~500 m between the Kaipo and John O’Groats may similarly be attributed to either the Waimea or Early Otira glaciation.

**S5.4.4 Waimaunga (MIS 8)**

Restoration of the Gorge River plateaux glacial deposits back to a Pyke valley source, restoration of the ~8 km bend in the Cascade valley, and synchronous alignment of all other major river valleys northwest and southeast of the Alpine Fault in South Westland, provide compelling evidence of the Waimaunga Glaciation (Figure 5.4). Considering all fault offset reconstructions over the last c. 500 kyrs, the landscape best fits to the time of this glaciation.
Elsewhere in Westland, Middle Pleistocene ice advances (MIS 10, 8, 6) are recorded only by patchy remnants of moraine and outwash terraces, but have been suggested to have been more extensive ice advances than those of the Late Pleistocene (e.g., MIS 4, 2) [Barrell, 2011]. More evidence is provided by weathered till assigned to MIS 8 near Edwards Pass in North Canterbury which is located ~20 km downvalley of MIS 2 termini [Rattenbury et al., 2006; Barrell, 2011]. As pointed out by Suggate [2004], each glacial advance tends to extend a shorter distance down-valley than the previous one, which may be an indication of progressive valley deepening and widening, such that a valley is able to accommodate a similar volume of ice over a shorter distance. Together with evidence introduced by this study, I suggest the Waimaunganga Glaciation was a major formative event in the sculpting and establishment of the modern river valleys in South Westland, and presumably elsewhere in the Southern Alps.

**S5.4.5 Nemona (MIS 10) and older**

Evidence for Nemona and older glaciations in South Westland is patchy. This study is unable to assign any glacial deposit to a specific glaciation older than MIS 8. Glacial deposit remnants on the Sara Hills, Hope Blue River Range, and Burmeister Tops to ~900 m elevation are poorly-preserved, lack surface morphology like moraine crests, and may potentially be attributed to Nemona or older glaciations (Figure 5.1C).

_Sutherland et al. [1995]_ interpreted the shallow-marine Teer Formation (with glacial dropstones) and C1 (oldest) glacial deposits at the Cascade Plateau to have been deposited c. 0.9 ± 0.4 Ma based on clasts they suggested were derived from the Pyke/Hollyford catchments and utilizing an Alpine Fault strike-slip rate of 27 mm/yr. From beach boulders and a section in the upper parts of the northeast branch of Carmichael Creek (see Figure 5.1C), they identified the C1 moraine lithologies as including Haast Schist, Red Mountain ultramafics, Maitai Group metavolcanics, Livingstone Sub-Group or Alabaster Group, Barrier Formation metaconglomerates and quartz-bearing plutonics. During the present study I examined outcrops and bedload in the eastern-most tributary of Carmichael Creek where it cuts the plateau and confirmed the lithologies described by _Sutherland et al. [1995]_ are also present as clasts there. However, it must be noted that all of these lithologies outcrop in the modern Cascade River catchment near Saddle Creek or elsewhere as examined in this study and as mapped by _Rattenbury et al. [2010]_. The clast provenance does not need to be explained by offset from the Pyke and Hollyford rivers and could be explained solely by evolution of the Cascade catchment (i.e., Maitai, Livingstone and Barrier rocks are exposed only on the
outside bend of a growing valley wall offset and were thus protected by subsequent glacial erosion). Furthermore, all Haast Schist samples examined exhibit textural zone IV textures which are common in the Cascade catchment but rare in the Pyke/Hollyford catchment. A lack of distinctive Darrans Suite plutonic rocks or Wooded Peak Limestone/marble in the morainal deposit is also inconsistent with a Pyke/Hollyford source. Although the age of the C1 moraine remains unresolved, the present study suggests it may be younger than previously suggested.

S5.5 Main Divide Capture

The Main Divide of the South Island has a relatively smooth trace from its junction with the Wairau section of the Alpine Fault in the north to the southernmost extent of Fiordland (Figure S5.3). The divide is symmetrical throughout the Fiordland orogen except near Milford Sound where it veers eastward to bound the south side of the Pyke/Hollyford catchment through two low saddles (with the south-flowing Eglinton and Greenstone catchments) and connect to the Southern Alps orogen. In contrast to Fiordland, the main divide through the Southern Alps is asymmetrical, with its position generally ~20 km southeast of the position of the Alpine Fault defining the steep western rangefront. However, there are three notable exceptions where the Main Divide is further from the Alpine Fault (~30–40 km): adjacent to the Haast/Landsborough, Arawhata, and Pyke/Hollyford catchments. Korup et al. [2005] calculated denudation and erosion rates in the 22 largest west-draining catchments in the Southern Alps from the Waitaha River in the north to the Cascade River in the south. They showed that hypsometric integrals (which quantify catchment topography) are scale dependent and highlight the extreme catchment area of the Arawhata (~750 km²) and Haast/Landsborough (~1300 km²) compared to the 20 other catchments examined (~50–450 km²). Although not examined in their study, the present study notes the 1150 km² area of the Pyke/Hollyford catchment. The present study suggests these three catchments stand well on their own because they were formerly east-draining catchments that have been captured by the West Coast rivers and have caused southeastward shifts of the main divide of 10–25 km. Restoring the main divide to its position prior to piracy of these three catchments yields a smooth drainage divide spanning the full length of Fiordland and the Southern Alps, which suggests there have been no other significant piracies of drainages that cross the main divide preserved since initiation of the modern Southern Alps/Fiordland orogen (Figure S5.3).
Modern Main Divide of the South Island forming the spine of the Southern Alps/Fiordland orogen overlain on a northwest-illuminated hillshade derived from 25 m LINZ DEM data. Notice the relatively straight trace of the Main Divide and the even size of catchments west of the divide, except in the vicinity of the Haast/Landsborough, Arawhata, and Pyke/Hollyford catchments. It is suggested large portions of these three catchments formerly drained to the east at which time the Main Divide would have had a smooth trace (see discussion in text).
S5.5.1 Haast/Landsborough

*Craw et al.* [2003] presented structural evidence to suggest a significant portion of the Haast/Landsborough catchment formerly drained east (Figure S5.3). A northeast (orogen-parallel) structural grain of kilometer-wide, fault-bounded slices of steeply-dipping, northeast-striking schist and greywacke controlled pre-Quaternary drainage of the Clark River south into the Makarora River and the Landsborough River south into the Hunter River. This trend is comparable to other south-draining rivers nearby (e.g., Hopkins River, Tasman River). *Craw et al.* [2003] suggest tectonic shortening in the northern part of the area combined with slip on a sinistral fault at one of the capture points explains sequential capture of the Clarke River and Landsborough River (and further the Burke and Wills rivers) by the west-flowing Haast River. The total capture resulted in an area of \( \sim 1000 \text{ km}^2 \) being transferred west of the Main Divide and local southeast migration of the Main Divide of \( \sim 25 \text{ km} \). The Haast/Landsborough catchment now flows across the regional structural grain and at present is one of the few west-draining catchments with significant orogen-parallel reaches in its headwaters.

*Cooper and Beck* [2009] examined a sodalite tinguaita and a fenite dredged from the bed of the Clutha River in Central Otago and concluded their only possible source is from the Burke River. The Burke River presently drains into the Haast/Landsborough, but the presence of these rocks indicates it formerly must have drained south into the Clutha catchment. Although poorly constrained in terms of age, both studies suggest a Late Pliocene-early Pleistocene (\( \geq 2 \text{ Ma} \)) capture of the Clarke and Landsborough rivers [*Craw et al.*, 2003; *Cooper and Beck*, 2009].

S5.5.2 Arawhata

The Arawhata is the third largest west-draining catchment in the Southern Alps. Although not examined in detail by the present study, it is suggested a similar capture event to the Haast-Landsborough may have occurred in the headwaters of the Arawhata River (Figure S5.3). The Joe River, Arawhata headwaters and Waipara River lie in roughly orogen-parallel \( \sim 1600 \text{ m} \) relief valleys that drain through narrow bedrock constrictions to join the wide northeast to north-trending Arawhata valley. Along the modern Main Divide, the Joe River valley has some of the steepest slopes anywhere in the Southern Alps. This study suggests the Joe/Arawhata headwaters may have formerly drained into the Dart River, while the Waipara River may have drained into the West Matukituki valley or potentially into the Joe. This
would indicate the main divide would formerly have transected the Olivine Ice Plateau (the largest ice plateau not on the South Island Main Divide) and extended over the Five Fingers, Waipara and Haast ranges to a position near the head of the Wilkin River. The high elevations of the saddles between the Joe and Dart (~1800 m) and between Waipara and Joe and West Matukituki (~1500 m) is comparable to the ~1900 m saddle between the Landsborough and Hunter rivers, which suggests these capture events may have occurred at similar times (c. ≥ 2 Ma).

**S5.5.3 Pyke/Hollyford**

Dextrally offset glacial deposits on the AUS plate can be used in part to unravel the evolution of the Pyke/Hollyford catchment, the second largest west-draining catchment in the Southern Alps at 1150 km$^2$. The modern Pyke River is underfit to the large size of its main valley and has a peculiar drainage pattern (Figure S5.4A). It is presently sourced from headwaters away from the Main Divide behind Big Bay. The river flows towards the sea through a narrow gorge-floored glacial valley, then takes an abrupt southward bend away from the Tasman Sea to flow through the 2 km-wide Pyke valley. The river has an exceptionally low gradient as it flows through the Pyke valley, spanning two large lakes (Wilmot and Alabaster) and dropping less than 15 m in elevation over 35 km of reach to its confluence with the Hollyford River. Below the confluence, the Hollyford River drops over 15 m in 9 km to join Lake McKerrow (elevation of 1 m). The Hollyford River then drains into the Tasman Sea at Martins Bay.

**Figure S5.4** Diagrams summarizing the proposed evolution of the Pyke & Hollyford valleys. (A) Modern drainage patterns in the vicinity of the Pyke valley overlain on a northwest-illuminated hillshade derived from 15 m LINZ DEM data. Notice the river in the main Pyke valley flows away from the Tasman Sea into the Hollyford River. The lack of significant alluvial fans at the Dry Awarua and Paulin Creek (where the modern Pyke is deflected by the Alpine Fault) suggest the modern Pyke headwaters may have flowed into Big Bay until very recently (e.g., post-LGM). (B) Drainage patterns at c. 300 ka (~8000 m of dextral offset restored. At this time the river in the main Pyke valley flowed from the main divide seaward to what is now the Gorge River area. The widespread glacial surfaces in the Gorge River area and lack of significant deposits associated with where the Lower Hollyford exits into the Big Bay area, suggests a drainage divide existed between the Lower Hollyford and Upper Hollyford/Pyke valleys at this time. Note the large amount of bedrock in the valley floor and generally narrower valley where the Lower Hollyford, Upper Hollyford and Pyke valleys meet. (C) Drainage patterns at some time before 300 ka, but younger than c. 2 Ma as constrained by topographic evolution in the Milford Sound region [Shuster et al., 2011]. The modern Hollyford valley presents the largest indenture in the otherwise smoothly asymmetric main divide of Fiordland and the Southern Alps; the highest peaks in Fiordland occur west of the Hollyford/Pyke valley junction, well off the modern main divide, while low ~500 m saddles currently form the main divide between the east-draining Eglington and Greenstone valleys and the west-draining Hollyford valley. It is proposed the main divide used to be through a position near the Hollyford/Pyke valley junction with the Upper Hollyford draining east through the Eglington valley.
The saddle between the Pyke River and the Dry Awarua River (which drains into Big Bay) has exceptionally low relief (< 15 m). The small, poorly-developed alluvial fan of Paulin Creek at this saddle is all that separates the drainages and suggests the Pyke headwaters may have flowed into Big Bay very recently (post-last glaciation, perhaps > 14 ka). Presumably dextral-normal movement across the Alpine Fault has deflected the Pyke headwaters into the Pyke valley away from the Tasman Sea. During peak glaciations, the flow through the Pyke valley is expected to have always been northward, away from the main divide and in line with the Upper Hollyhord Valley. Glaciated bedrock between Big Bay and the Pyke River indicate ice has previously flowed over the entire 7 km-wide gap towards Big Bay. An offset hillslope of ~1450 m (Figure 5.9) on the south side of the Pyke/Big Bay gap, indicates this was the most recent time (c. 55 ka) when ice flowed from the main Pyke valley into Big Bay (it is further suggested the well-preserved lateral moraines at Awarua Point on the north side of Big Bay date to this time).

As described previously, the Waimaunga Glaciation (~8 km of offset restored, c. 300 kyr) was a major formative event in sculpting drainages and transferring and depositing large masses of sediment to terminal and lateral moraines. At this time a glacier flowed north from the Upper Hollyhord valley through the Pyke valley to deposit the widespread glacial surfaces now present in the Gorge River area (Figure S5.4B; Figure 5.6). Because of the wide piedmont glacier present in the Gorge River area at this time, and the comparably small outlet for the Lower Hollyhord valley to Big Bay and lack of related glacial deposits, it is argued a drainage divide existed between the Upper Hollyhord/Pyke and Lower Hollyhord at this time. Note that the valleys are particularly narrow at the junction of the Pyke, Upper Hollyhord, and Lower Hollyhord, and that glacial-sculpted bedrock extends a considerable distance into the floor of these valleys to form a bedrock channel within the valley bottom that is only ~600 m wide at the junction.

Even without the Lower Hollyhord, the Upper Hollyhord/Pyke catchment is still the second largest northwest-draining catchment in the Southern Alps. The Pyke/Hollyhord catchment size, position of the highest peaks in Fiordland off the Main Divide near the Pyke/Hollyhord confluence, and narrow bedrock floored valley at the Pyke/Hollyhord confluence are all used to argue that the Main Divide was formerly at this confluence (Figure S5.4C). The Pyke valley and Lower Hollyhord valley would have drained north, but the Upper Hollyhord would have drained south through what is presently a low ~500 m divide and into the south-draining Eglinton valley. This would imply the subsequent piracy captured > 450 km² of catchment...
area to the west side of the Main Divide and shifted the position of the Main Divide ∼30 km
to the south. If the Main Divide is placed through a position near where the Cleddau River
enters Milford Sound, the result would be a smooth Main Divide linking the highest peaks of
Fiordland (Mt Tutoko, Mt Madeline) and the Southern Alps. Shuster et al. [2011] used (U-
Th)/He thermochronology at Milford Sound to reveal that glacial erosion during the past c. 2
Myr removed the entire pre-Pleistocene landscape by headward propagation of trunk valleys
toward high elevation plateaus. If the Upper Hollyford Main Divide piracy described
occurred after propagation of the Cleddau valley, this would be consistent with a smooth
Main Divide trace at c. 2 Ma.

While the mechanism for this Main Divide piracy is unknown, local faulting, landsliding and
 glaciations may all have been contributing factors. The west-directed convergence of the PAC
plate is expected to drive the main divide westward and lead to headward erosion and capture
by West Coast rivers. The Lower Hollyford, Pyke, Upper Hollyford and Eglinton valleys all
follow major pre-Eocene regional faults (Glade-Darrans Fault, Hollyford Fault System,
Figure 2.1B; Figure S2.11A). While it is clear significant differential uplift has occurred across
the Glade-Darrans Fault (Supplement 2), it is unknown whether this structure remains active
today. However, the geomorphic expression of the fault on the western side of the Hollyford
valley, together with friable rock and soft clay gouge in its fault zone [e.g., Sutherland, 1995],
suggest the Glade-Darrans Fault may presently be active.
Appendix

View looking northeast at a major fault (diagonal at center) and Pacific plate fault-damaged rocks within the Alpine Fault zone at the Jerry River (foreground). Total height of this outcrop is ~150 m.
### A1. Correlation of University of Otago (OU) and NCB Field (N) Numbers

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TS = Thin Section  MS = Mineral Separate  HS = Hand Sample
A2. Conferences and Conference Abstracts Resulting from Thesis Research

First Authored (presented):


Co-Authored:


Full Abstracts (as above):


Aerial light detection and ranging (LiDAR) data were collected from a 1.5 x 34km swath encompassing the central Alpine Fault (the major plate-boundary structure on the South Island of New Zealand). By incorporating interpretations of the new LiDAR data with aerial photo interpretation, subsurface fault geometry from the recent Deep Fault Drilling Project (DFDP) boreholes, and previous geologic mapping, we explore the significance of several orders of magnitude of transpressional partitioning observed on the central Alpine Fault at 1-10 m scales.

We recognize 9 first-order (1-10km scale) thrust and strike-slip serial partitions in the LiDAR coverage area, which mostly accord with segments mapped previously. The overall strike of thrust segments average 047 and strike-slip segments 082. Thrust partitions tend to be longer. This partitioning results from stress perturbations arising from hanging wall topographic variations in a transpressional regime.

Over 250 second-order dextral-thrust, dextral and dextral-normal fault traces were identified in the LiDAR data. Mean trace length is 260m with a few traces longer than 1km. Many of these traces have scarp heights to 20m indicating individual scarps represent the culmination of several surface rupture events, which have a characteristic heave < 1.5m. Fault kinematics indicate steeper-dipping, more E-W-striking faults slip at lower differential stress than shallow-dipping ones in the modern stress regime. We propose the partitioned structures reflect the system forming fault orientations that are favorable for shear.

At the fault trace scale, parallel partitioning and en echelon partitions are more common than serial partitions. We use constraints from DFDP boreholes to suggest this second-order geometry is related to the depth to basement in the footwall.

We propose the surface geometry reflects subsurface structure comprising an asymmetric positive flower structure bounded by dextral-normal and thrust faults rooted on a planar, moderately southeast-dipping, dextral-reverse fault plane at shallow depths (< 600m). We map 100 pressure ridges, 0-15° counter-clockwise oblique to the nearest fault trace strike, consistent with transpression.

The near-surface complexity in this region has formed despite paleoseismic evidence that the Alpine Fault ruptures during one through-going ~M8 event every 100-300 years. The complexity
therefore illustrates that these large ruptures rooted in basement, do not remain localized when propagating through unconsolidated surficial sediment under low confining pressures. This region also highlights the difficulties involved in determination of surficial slip rates in partitioned fault zones.

(2)  Same as (1).


New Zealand’s Alpine Fault is mostly a moderately SE-dipping dextral reverse plate boundary structure, but at its southern end, strike-slip-normal motion is indicated by offset of recent surfaces, juxtaposition of sediments, and both brittle and ductile shear sense indicators. At the location of uplift polarity reversal fault rocks exhumed from both the hangingwall Pacific and footwall Australian Plates are juxtaposed, offering a remarkably complete cross section of the plate boundary at shallow crustal levels. We describe Alpine Fault damage zone and fault core structures overprinted on Pacific and Australian plate mylonites of a variety of compositions, in a fault-strike perpendicular composite section spanning the reversal in dip-slip polarity.

The damage zone is asymmetric; on the Australian Plate 160m of quartzose paragneiss-derived mylonites are overprinted by brittle faults and fractures that increase in density towards the principal slip surface (PSS). This damage zone fabric consists of 1-10m-spaced, moderately to steeply-dipping, 1-20cm-thick gouge-filled faults, overprinted on and sub-parallel to a mylonitic foliation sub-parallel to the PSS. On the Pacific Plate, only 40m of the 330m section of volcaniclastic-derived mylonites have brittle damage in the form of unhealed fractures and faults, as well as a pervasive greenschist facies hydrothermal alteration absent in the footwall. These damage-related structures comprise a network of small-offset faults and fractures with increasing density and intensity towards the PSS.

The active Pacific Plate fault core is composed of ~1 m of cataclasite grading into folded protocataclasite that is less folded and fractured with increasing distance from the PSS. The active Australian Plate fault core is < 1.5m wide and consists of 3 distinct foliated clay gouges, as well as a 4cm thick brittle ultracataclasite immediately adjacent to the active PSS. The Australian Plate foliated clay gouge contains stringers of quartz that become less continuous and more sigmoidal toward the PSS, indicating a strain gradient across the gouge zone. Gouge textures are consistent with deformation by pressure solution. Intact wafers from one of the gouges, experimentally sheared in a biaxial configuration under true-triaxial loading at $\sigma_n^*$ = 31MPa and $P_t$ = 10MPa, yielded a friction coefficient, $\mu_{ss} = 0.32$ and displayed velocity strengthening behavior. No significant re-strengthening was observed during hold periods of slide-hold tests.

Well-cemented glacial till (c. 8000 years old) which caps many outcrops, is a marker that shows that the damage zone is not active in the near surface, but most of the fault core is. The active near-surface damage zone here is < 40m wide and the active fault core is < 2.5m wide. Both overprint a much wider, inactive damage zone. The combination of rheologically-weak Australian plate fault rocks with surface rupture traces indicates distinctly different coseismic and interseismic behaviors along the southern strike-slip-normal segment of the Alpine Fault.

The Alpine Fault is best known as a moderately SE-dipping dextral reverse plate boundary structure, but along its southernmost 70km onshore, strike-slip-normal motion is indicated by offset of recent surfaces, juxtaposition of sediments, and both brittle and ductile shear sense indicators. At the location of uplift polarity reversal, fault rocks exhumed from both the Pacific and Australian Plates are juxtaposed, offering a remarkably complete cross section of the plate boundary at shallow crustal levels.

The damage zone is asymmetric; on the Australian Plate 160m of quartzose paragneiss-derived mylonites are overprinted by brittle faults and fractures that increase in density towards the principal slip surface (PSS). On the Pacific Plate, only 40m of the 330m section of volcanoclastic-derived mylonites have brittle damage and pervasive hydrothermal alteration.

The active Pacific Plate fault core is composed of ~1 m of cataclasite. The active Australian Plate fault core is < 1.5 m-wide and consists of foliated clay gouges with pressure solution textures and stringers of quartz that become less continuous and more sigmoidal toward the PSS. Intact wafers from an Australian Plate gouge, experimentally-sheared in a biaxial configuration under true-triaxial loading at $\sigma_n = 31\text{MPa}$ and $P_t = 10\text{MPa}$, yielded a friction coefficient, $\mu_{ss} = 0.32$ and displayed velocity strengthening behavior. No significant re-strengthening was observed during hold periods of slide-hold tests.

Well-cemented glacial till (~8000 years old), which caps many outcrops, is a marker that shows that the damage zone is not active in the near-surface, but most of the fault core is. The active near-surface damage zone here is < 40m wide and the active fault core is < 2.5m wide. Both overprint a much wider, inactive damage zone. The combination of rheologically-weak Australian Plate fault rocks with surface rupture traces indicates distinctly different coseismic and interseismic behaviors along the southern strike-slip-normal segment of the Alpine Fault.


A large proportion of coseismic slip is thought to occur along narrow principal slip zones (PSZs) that are millimeters to centimeters- thick. Understanding the physical properties of PSZ gouges and cataclasites, and their along strike extent, is a primary goal of fieldwork being done in conjunction with the Deep Fault Drilling Program (DFDP), Alpine Fault. Core retrieved from two shallow boreholes at Gaunt Creek contains mylonites with a Pacific Plate Alpine Schist protolith, cataclasites derived from Pacific Plate mylonites and from Australian Plate felsic igneous rocks, gneisses, and possibly metasediments, and three PSZ gouges. These three PSZ gouges are: (1) DFDP-1A PSZ gouge (90.5 – 90.8 m depth), which occurs at the contact between Pacific Plate cataclase and Australian Plate Late Quaternary gravel; (2) DFDP-1B PSZ1 gouge (128.0 – 128.3 m depth), which occurs below Pacific Plate cataclasite and above mixed Pacific and Australian Plate-derived cataclasite; (3) DFDP-1b PSZ2 gouge (143.83 – 143.94 m depth), which occurs below the mixed protolith cataclasite and above Australian Plate augen mylonite.

Microstructurally, PSZ gouges in the core are similar to those documented in a nearby outcrop at Gaunt Creek by Boulton et al. (2012). Brown PSZ gouge layers contain reworked fault gouge clasts together with subrounded to rounded clasts of mylonite, ultramyloinitc, quartz and carbonate-rich fragments. From quantitative XRD, the brown PSZ gouges in DFDP-1b comprise 18% dioctahedral smectite, along with quartz (25 – 29%), orthoclase/microcline (3 – 4%), albite (24 - 25%), calcite (8 - 9%), kaolinite (0 - 6%), muscovite/illite (12 - 17%), chlorite (0 – 1%), and trace amounts of pyrite. Boulton et al. (2012) reported similar mineralogy for Gaunt Creek outcrop and Waikukupa Thrust outcrop PSZ gouges and found that the PSZ gouges have lower friction coefficients and lower permeability than surrounding cataclasites. Mineralogically,
microstructurally, and geochemically similar brown PSZ gouges also occur at localities north and south of Gaunt Creek, from Little Man River to Robinson’s Creek, an along strike distance of 120 km. Everywhere mapped, brown PSZ gouges form at the contact between Pacific Plate and Australian Plate-derived cataclasites, which, importantly, do not contain smectite. Smectite-bearing gouges are generally absent on shallow dipping dextral-reverse faults at the toes of large thrust sheets, where plate boundary cataclasites overlie Late Quaternary gravels in sharp contact. Our results suggest that PSZ gouges retrieved in the DFDP-1 cores are commonly present on moderately dipping (average orientation 043°/30°SE; Norris and Cooper, 2007) dextral-reverse faults along the central Alpine Fault, and we discuss modes of PSZ gouge formation.

References


Integrated field, mineralogical, microstructural and rock deformation studies show that the Alpine Fault core comprises frictionally weak, clay-rich fault gouges (e.g., Boulton et al., 2012). Along the central section of the Alpine Fault, oblique thrust segment principal slip zones (PSZs) are composed of < 0.05 m-thick striated fault gouges composed of quartz-albite/anorthite-orthoclase/microcline-calcite-smectite-muscovite/illite-chlorite. Friction experiments on saturated PSZ gouges show that these fault rocks are velocity strengthening, with friction coefficients between 0.31 and 0.44. In contrast, clast-supported foliated and nonfoliated cataclasites in the immediate hanging wall are sometimes velocity weakening, with friction coefficients between 0.51 and 0.57.

The southern onshore section of the Alpine Fault is lithologically and structurally distinct (Barth et al., submitted). Here, clay-rich gouges form 1 m to 12 m-thick subvertical fault cores. Mineralogy of the gouges varies along strike, with a 1.5 m-thick fault core at the Martyr River containing velocity strengthening quartz-orthoclase/microcline-albite-calcite-muscovite/illite-chlorite gouges with friction coefficients between 0.32 and 0.37. At McKenzie and Hokuri Creeks, a wider fault core contains saponite-chlorite-serpentinite-rich weaker velocity strengthening gouges with friction coefficients between 0.11 and 0.13.

Clay transformation reactions produce weak PSZ gouges on the Alpine Fault. The presence of recycled gouge clasts suggests that earthquake ruptures preferentially reshear these clay-rich fault rocks. Velocity strengthening materials should theoretically arrest earthquake rupture propagation, but positive feedback mechanisms activated during high velocity slip may weaken the principal slip surface and facilitate large magnitude (~Mw 8), large displacement (~8 – 9 m horizontal slip) earthquakes on the Alpine Fault.

References


A light detection and ranging (LiDAR) survey was flown along the central Alpine Fault between Franz Josef and Whataroa to uncover fault-related surface deformation obscured by dense bush. The subsequent 2-m DEM produced beautiful bare earth images of fault expression along the rangefront. Initial results from this work include exposing a scarp and thrust fault that probably moved during the AD 1717 earthquake (De Pascale & Langridge, 2012) and detailed analysis of the orientation of the fault and its shallow structure (Barth et al., 2012). A goal of the LiDAR data has been to improve our understanding of the fault location and structure near the two DFDP sites (Gaunt Creek, Whataroa valley). Near the DFDP-1 site at Gaunt Creek, the LiDAR revealed a NNW-striking thrust trace and a flight of uplifted terraces on the hangingwall side of the fault. In addition, the partitioned nature of the surface traces of the fault was recognised SW of Gaunt Creek, where single- and multi-event displacements are observed along the inboard zone of strike-parallel slip. While surface traces are absent where the Alpine Fault crosses the Whataroa river alluvial fan, the LiDAR data constrains fault structure to the W and NE, which in turn constrains the fault geometry beneath the proposed DFDP-2 site to the SE. NE of the river, the Alpine Fault deforms a post-glacial age fluvial surface. In this area particularly (Parker to Vine creeks), the shallow partitioning and dip of the fault and the style of deformation can be elucidated from the shape of the deformed terrace using profiles from the LiDAR DEM. West of the Whataroa River, several important strike-slip traces offset streams incised into a similar fluvial surface, which we have dated at c. 11 kyr. This talk presents results of mapping and analysis in terms of deformation caused by the fault to Holocene landforms.


Some proportion of the elastic strain energy stored in the crust is dissipated on and around fault planes both co- and aseismically, rather than converted to radiated energy. We have examined outcrop to microscale fabrics of active natural fault rocks from principal slip zones (psz; defined as the zones that accommodated most shear displacement) in various tectonic settings, and obtained laboratory data about frictional behaviour and microstructural evolution. Here we consider how observed microfabrics broadly reflect the dissipative processes that operated within these psz.

1. New Zealand’s Alpine Fault is capable of generating $M_w \sim 8$ earthquakes with a return period of 200-300 years and is near the end of a major earthquake cycle. Continuous core of the fault psz recently recovered from its locked central section demonstrates most coseismic slip was localised to a cm-thick layer of highly comminuted ultracataclasite derived from a quartzofeldspathic protolith. This layer contains recycled gouge clasts illustrating it remains the locus of slip through multiple earthquakes, despite displaying velocity-strengthening behaviour in laboratory tests. It is surrounded by more permeable but cemented cataclasites into which psz gouge has been injected, suggesting thermal pressurisation occurred. The main dissipative mechanisms in this environment are inferred to be particle fragmentation and frictional heating resulting in thermal pressurisation.
of free fluids, both accompanying seismic slip. However, we recognise fragmentation probably ceased with the advent of thermal pressurisation, and that it is generally unlikely a significant proportion of available elastic strain energy is dissipated on-fault by these mechanisms.

2. Elsewhere, the Alpine Fault’s psz is metres thick and composed of a gouge of foliated clay minerals (including Mg-rich smectites). This difference in structural style reflects protolith mineralogy since in this region the fault cuts a mixture of quartzofeldspathic and ultramafic rocks. Foliated and folded packages of gouge are common and microstructural evidence strongly suggests solution-transfer is an important deformation mechanism in these gouges, which are frictionally weak and velocity strengthening in experiments. Some proportion of energy dissipation on this part of the fault therefore occurs aseismically, although there is also evidence for some coseismic energy release.

3. The microfabric of natural fault gouge from the creeping segment of the San Andreas Fault, (‘SDZ’ sample from the SAFOD borehole) comprises packages of foliated and folded clay minerals (including Mg-rich smectites). Others have inferred this microstructure is characteristic of material that deforms by continuous creep. This is consistent with our experimental observations at intermediate slip velocities, where disaggregated gouge re-develops a microstructure comprising foliated and folded packages of clay minerals when shear is distributed throughout the gouge layer. At faster experimental slip rates, localised through-going slip surfaces and hydrofractures form, indicating increasing importance of thermal pressurisation and rock fragmentation as coseismic dissipative mechanisms.

4. The Tohoku plate boundary interface subduction thrust, which we recently sampled during IODP Expedition 343, also comprises > 1m thickness of foliated, sheared clay with a variably intense scaly fabric and occasional microfolds. We infer this material would have accommodated continuous creep under strain hardening conditions. Embedded within it are rare, through going, zero thickness planar surfaces along which we infer strain localised to accommodate rapid loading (possibly including earthquake-rate slip). Elsewhere in the core, we observed structures more likely to result from recent coseismic slip. These are localised (mm-thickness) zones of ultracomminuted material surrounded by fractured breccias. In these zones both rock fragmentation and frictional heating were important on-fault dissipative processes.

To summarise, natural faults display a range of microfabrics within psz that may be localised or distributed through metre thicknesses. Creeping faults generate distributed zones of foliated and folded material, whereas faults that slipped seismically generate more localised zones of comminuted material, and dissipate some energy coseismically via fragmentation. Finally, there is evidence frictional heat was generated coseismically on most of the localised faults, but we have not observed pseudotachylytes so infer this heat is dissipated into fluids or by mineral decomposition reactions.

On the South Island of New Zealand, the Alpine Fault strikes continuously for over 800 km with no along strike separations greater than 5 km. Unlike other major active continental strike slip faults such as the North Anatolian Fault in Turkey and the San Andreas Fault in California, there have been no known surface ruptures on the Alpine Fault in the c. 170 year historic period; there are also no aseismic creeping sections on the Alpine Fault (e.g., Sutherland et al., 2007 and references therein). Integrated field, mineralogical, microstructural and rock deformation studies, however, show that the Alpine Fault core comprises frictionally weak, clay-rich fault gouges (e.g., Warr and Cox, 2001; Boulton et al., 2012).

Along the central section of the Alpine Fault, for a minimum of 115 km from Gaunt Creek to Robinson’s Creek, thrust segment PSZ’s are composed of 1 cm to 50 cm-thick striated fault gouges composed of quartz-albite/anorthite-orthoclase/microcline-calcite-smectite-muscovite/illite-chlorite. Friction experiments on saturated fault core gouges show that these fault rocks are velocity strengthening, with friction coefficients between 0.31 and 0.44. In contrast, clast-supported foliated and nonfoliated cataclasites in the immediate hanging wall are sometimes velocity weakening, with friction coefficients between 0.51 and 0.57.

The southern onshore section of the Alpine Fault is lithologically and structurally distinct (Barth et al., submitted). Here, clay-rich gouges form a 1 m to 12 m-thick subvertical fault core. Mineralogy of the gouges varies along strike, with a 1.5 m-thick fault core at the Martyr River containing quartz-orthoclase/microcline-albite-calcite-muscovite/illite-chlorite gouges with friction coefficients between 0.32 and 0.37 and velocity-strengthening behavior. Wider fault cores contain saponite-chlorite-serpentinite-rich clay gouges and are also velocity strengthening, with friction coefficients between 0.11 and 0.13.

The Alpine Fault is a long-lived crustal structure; evidence from outcrop cross-sections (described above), as well as core logs recovered from the Deep Fault Drilling Program (DFDP) extend observations made by Warr and Cox (2001) that clay transformations are important weakening mechanisms on both the central and southern Alpine Fault. Moreover, the presence of recycled gouge clasts in all studied fault core gouges suggests that earthquake ruptures preferentially reshare these clay-rich fault rocks. Although velocity strengthening materials should theoretically arrest earthquake rupture propagation, positive feedback mechanisms activated during high velocity slip such as thermal pressurization and perhaps endothermic clay dehydration reactions may dramatically weaken the principal slip surface and facilitate large magnitude (~ Mw 8), large displacement (~ 8 – 9 m horizontal slip) earthquakes on the Alpine Fault.
The Alpine Fault accommodates Pacific-Australian plate boundary convergence on a single northeast-southwest striking structure with a surface trace at least 800 km long and a cumulative offset of ~470 km. In this study, we sampled principal slip zone (PSZ) fault gouges at 5 locations on the Alpine Fault: Gaunt Creek, Waikukupa River, Fault Creek, Core Creek and Hokuri Creek, a NE to SW along strike distance of ~250 km. Using multiple analytical techniques, we measured along strike variations in permeability, thickness, frictional strength and mineralogy. All the fault gouges have very low permeability ($k$) between $k = 1.2 \times 10^{-19}$ m$^2$ and $k = 3.6 \times 10^{-21}$ m$^2$. From NE to SW, we observed a systematic two order of magnitude increase in fault gouge thickness from 10 cm at Gaunt Creek to 11 m at Hokuri Creek. This increase in gouge thickness correlates directly with a decrease in the friction coefficient ($\mu$) from $\mu = 0.44$ at Gaunt Creek to $\mu = 0.11$ at Hokuri Creek. The Alpine Fault gouges markedly weaken where the fault core widens to ~1.5 m at Fault Creek. Here, the orientation and kinematic shear sense of the Alpine Fault displays an abrupt transition. To the northeast, the plate boundary structure is moderately SE-dipping and accommodates dextral reverse slip; to the southwest, it is very steeply SE-dipping with dextral normal motion. XRD and SEM-EDS data show that the stronger PSZ gouges contain quartz-chlorite-muscovite-feldspar minerals; weaker PSZ gouges contain quartz-clinochlore-muscovite-montmorillonite mineral assemblages with well-developed phyllosilicate lamellae. At Hokuri Creek, the weakest gouge contains lizardite-clinochlore-talc-saponite. We suggest that the increase in PSZ thickness and corresponding decrease in friction coefficient SW of Fault Creek may contribute to the observed reversal in dip-slip polarity, promote aseismic fault creep, and affect the nature of earthquake rupture nucleation and propagation along this segment of the Alpine Fault.

In the central South Island, the NE-striking, dextral-reverse Alpine Fault forms the principal component of the Australia-Pacific plate boundary. The fault accommodates high slip rates of ~25-29 mm/yr (dextral) and up to 6-11 mm/yr (reverse), mostly uplifting Pacific plate rocks that form the Southern Alps. However, the associated rapid uplift and erosion and dense temperate rainforest along the western side of the island have typically hampered geological efforts to better understand the surficial structure of the Alpine Fault.

A 34 km long x 1.5 km wide swath of airborne LiDAR (Light Detection and Ranging) survey along the central section of the Alpine Fault between Franz Josef and Whataroa has been flown and processed. A 2-m DEM developed from the LiDAR has provided unhindered bare-earth images of the tectonic geomorphology along the fault.

Two of the research highlights for this work have been (i) the mapping of more than 250 NE- to ENE-striking dextral- to dextral-normal slip fault traces and N- to NNE-striking dextral-reverse to thrust fault traces; and (ii) paleoseismic trenching of a mapped linear scarp on the true right side of Gaunt Creek adjacent to the DFDP-1 drill site. The former confirms the models of partitioning and shallow fault segmentation shown by Norris and Cooper (1995) but delineates more complex, finer scale structure. The latter documents repeated late Holocene, low-angle faulting across a NNW-striking fault scarp. We have documented faulting on this trace which corresponds with the most recent rupture event on the fault in c. AD 1717.


In central South Island, New Zealand, the dextral-reverse Alpine fault forms the principal component of the Australia-Pacific plate boundary. The fault typically accommodates slip rates of the order of ~27-29 mm/yr (dextral) and up to 6-11 mm/yr (reverse), mostly uplifting Pacific plate rocks that form the Southern Alps. However, the associated high relief, rapid uplift and erosion and high rainfall and accompanying dense temperate rainforest along the western side of the island has typically hampered geological efforts to better understand the neotectonics of the Alpine fault. LiDAR data have been acquired over a 34 km stretch of the fault between Whataroa in the northeast and Franz Josef in the southwest to test the viability of this technique under dense vegetation and in steep, dissected terrain. LiDAR has been collected from a fixed wing base (1300m above ground level) at a frequency of 70k Hz, with 33.5 Hz scan frequency and a 39° field of view. We employed a strategy of flying a dense pattern of 6 flight lines across a swath width of 1.3 km. This creates areas of both single and double overlap coverage that have allowed for accurate landscape models to be created. Results show that this strategy has provided an optimum level of forest penetration and ground returns. Initial results show remarkable level of detail in DEM's of the landscape along the Alpine fault. Examples of results presented here include: Franz Josef, where the fault traverses the township; and Gaunt Creek, where a Deep Fault Drilling Project will be sited in early 2011.
A3. Fault Rock Preparation Method for Thin Sections

Nicolas Barth and Brent Pooley – March 2012

Field Sampling:

1. Use paint-based markers or correction fluid for best preservation of orientation. If sample is wet, wax-based China markers are best bet. Very carefully remove sample from outcrop using a rock hammer, pry bar or brick chisel as necessary. Clay-rich gouges often require a brick chisel driven into a side of a sample with the material removed on the side of the chisel opposite the sample and this step repeated for all the other sides. Clay-rich gouge orientation is best marked in the field by scribing into the gouge or better yet pressing small twigs into the gouge. Immediately wrap in aluminum foil (to preserve integrity) and place in a zip-lock bag (to preserve moisture).

Gluing/Cutting/Mounting:

2. Re-mark orientation with a paint-based marker if possible (permanent markers and China markers do not survive the gluing process very well). Heat briefly (~5 minutes) on a hot plate at medium heat (remove aluminum foil if sample is intact enough). Having aluminum foil between the sample and hot plate makes the sample easier to remove once the glue has hardened. Coat the entire sample with Araldite by “painting” it on with a nail or other implement. This can be very time-consuming, but it is important that the sample is 100% covered in a thin coat. Leave to harden for a minimum of 20 minutes (often 1-2 hours is necessary).

3. Once glue has hardened, check to see whether the sample is thoroughly coated in glue. If you were unable to glue a portion of the sample, rotate the sample and finish gluing as above.

4. Once completely coated, reorient sample and mark the desired cuts. Use a thin diamond saw with minimal water (no water for clay-rich gouges) to make a cut as smoothly as possible. Immediately remove from the saw once cut. If the sample happens to be intact enough, several cuts can be made at once. However, the more unglued faces, the significantly weaker the sample! Clay-rich gouges will only survive with one unglued surface. The cut surface will probably be coated in particles and a clay residue that needs to be very carefully removed. Discretion is needed here based on the competency of the sample. I found the best method of cleaning the cut surface is to place the sample cut face down underneath a gentle stream of water from a faucet. This minimizes erosion as the water hits the top of the sample and gently rolls across the underside of the sample. Gently rock the sample back-and-forth and adjust the water flow for best results. If the sample is intact or homogeneous enough, using a finger to gently rub away the residue can be effective.

5. Once the cut surface is clean, place the sample cut face up on the hot plate. Once dry, glue as per Step 2.

6. Repeat cuts as per Step 4 and glue as per Step 2 until you have a billet. Carefully label sample name and orientation on the back of the sample.

7. Glue the surface to be mounted on a glass slide. Once the glue is hard, place billet face down on the hot plate. Remove after about a minute and scrap the excess glue from the surface to be mounted using a glass slide (this saves significant effort in the polishing process, care is needed with clay-rich samples).

8. Dry sand the sample using 180, 600 and 1000 grit sandpaper on a very flat clean surface (preferably glass). This is also time-consuming. Preferably use a different sheet of sandpaper for each sample to avoid contamination and be sure they are clean of fine grains that could scratch the sample. Sand the sample with 180 grit sandpaper until relatively flat but with a very thin layer of glue remaining. At this point, move on to 600
grit (180 tends to be too rough on uncoated surfaces). Sanding the samples in one direction (instead of back-and-forth or circular motions) tends to minimize “plucking” of grains. However, much care is needed to apply force to the sample evenly as it is very easy to end up with rounded edges on the sample. If this happens, go back to a coarser grit. After using the 1000 sandpaper, the sample should have a slight polish (especially on any quartz clasts) and only very minor scratches. Check that the surface is perfectly flat using a glass slide or the billet will not mount properly.

9. Mount billet to glass slide using Araldite as you would with any other sample and place in an oven to harden. Check the slide after 2 minutes to make sure bubbles have not formed, and again after 10 minutes. After 30 minutes the glue should have hardened. Remove sample from the oven and allow plenty of time to cool before continuing (slides can crack easily due to thermal expansion of clays).

10. Use a diamond scratcher to carefully scratch the name and orientation onto the back of the slide. Carefully cut the billet from the slide using a thin-bladed diamond trim saw (and minimal water) with a jig set to cut about 2mm away from the slide. Clay-rich samples will then need to be carefully washed or cleaned with compressed air to avoid erosion as in Step 4, and then re-glued.

11. Very carefully grind the sample down on grinding wheels using 100, then 240 grit. If clays start to erode, re-glue before continuing. The samples will go very quick so be careful! Next use 400 (then 600) grit on a wet glass sheet until the sample is very slightly thicker than 30 microns. Clay-rich samples: Dry sand only at 180, 600 and 1000 grit as per Step 8.

12. At this point it is very important to coat the surface with Araldite using a finger to rub evenly across the sample, and then wipe most of it off with a quality paper towel (one that will not shed fibers). Place in an oven until the glue hardens.

13. Place “button” on hot plate and melt an even droplet of wax on its top surface. Immediately, heat the thin section on the hot plate for 3-4 seconds, lightly press the thin section onto the button, remove the button and thin section from the hot plate and use a wooden dowel to press lightly on the thin section so that it is evenly on the button. Let sit for several minutes until the sample cools. Be very careful not to have buttons or thin sections too hot at this stage (or at very different temperatures at time of contact). Clay-rich samples break very easily at this stage due to thermal expansion!

14. Sand the thin section using 1000 grit sandpaper. Again sand only in one direction! Hold sample so that a corner faces in the direction of sanding and give the sample 3-4 passes before rotating 90° and repeating for the 3 remaining corners. Then very lightly sand the sample in a back-and-forth motion for about 20 seconds.

Polishing:

15. The remaining steps utilize a flat lap polisher. Clean the sample with water (compressed air only for clay-rich samples!). Place sample on a pad with 3 micron diamonds with the heavy weight holding it in place. Run for 20 minutes at a low RPM.

16. Clean sample with water (compressed air only for clay-rich samples!). Use a separate pad with 1 micron diamonds with the lighter weight holding it in place. Run for 10 minutes at a moderate RPM. Clay-rich samples are done at this point!

17. Clean sample with water. Use a black rubber-like pad with colloidal silica with the lighter weight holding it in place. Run for 4 minutes at a moderate RPM. Use a squirt bottle to re-lubricate the pad with colloidal silica every minute or so.

18. Clean sample with water. Place sample button down on hot plate. Remove the thin section from button as soon as wax melts. Let cool. Once cooled, wet the thin section and scrape off wax using a razor blade.
A4. Petrography of \(^{40}\text{Ar}/^{39}\text{Ar}\) Dated, Musc/Amph Samples

Refer to Chapter 2 and Supplement 2 for analytical procedures and results of \(^{40}\text{Ar}/^{39}\text{Ar}\) analyses.

**OU82942/N090610A**: sil grade quartzo-pelitic paragneiss with qtz+fsp+musc leucosomes (Greenland Group)

**Hand**: qtz+bio dominant, moderately strong foliation defined by bio+musc, contains boudinaged foliation parallel, but often isoclinally folded, coarse grained qtz+fsp+musc leucosomes (not sampled for dating)

**Thin**: qtz (55%), chl (15%), musc (10%), fsp (Incl. Kfs) (10%), ttn (8%) (almost exclusively enclosed within chl, associated with lenticular K-spar, suggesting the three phases represent retrogressed biotites), ap (2%), bio as fresh grains preserved as inclusions in qtz (trace).

Qtz displays dominantly subgrain rotation recrystallization (SR) textures (sweeping undulose ext, deformation lamellae, subgrains, relatively straight grain boundaries, similar grain/subgrain sizes). Typical grain size is 0.5 x 1.5 mm. Very occasional qtz def bands cross cutting fabric.

Chl pseudomorphs after bio and musc both present in foliation. Both seem stable, particularly musc. Typical musc size is 0.2 x 3 mm. Occasional more complex musc xtls 1.5 x 2.5 mm. Ser as random f.g. flakes within fsp.

**OU82946/N110603A**: hbl+fsp+qtz L>S dioritic orthomylonite (Pac plate, hosted in Maitai-derived, chl-ep-qtz mylonites)

**Hand**: hbl-rich, L > S mylonite. Some hbl have 6:1 aspect ratios. Larger grains less altered with more mylonitic tails. Some late stage shears+veins.
Thin: alt fsp (originally pl) (50%), hbl (40%), qtz (10%)

Fsp much finer grained and altered than hbl. Hbl behaving more ductily as fsp grains are more equant. Qtz is uncommon except in late veins but has undulose ext.

Hbl commonly fractured along cleavage with some alteration. Inclusions are also common. Some clean grains too. Hbl commonly has prism-like subgrains with continuous sweeping undulose extinction. Myl tails and bending of cleavage into fish shapes common. Grains vary in size from 1 x 3 mm to 1 x 4 cm.

Coarse porphyroblasts/porphyroclasts of hbl, fractured and disrupted with fracture planes composed of pale trem/act + chl + ttn, and with fibrous overgrowths of trem/act parallel to prism axis/mylonitic foliation. Rare lenticular patches of ribbon qtz, but main matrix is composed of very fine aggregate of ep with interstitial fsp (alb?) probably pseudomorphs of original Ca-pl. Cut by vein composed of non-mylonitic qtz and cal. Relationships suggest chl-ep alteration occurred at a similar time to the mylonitization (and at the same time as the host Maitai-derived volcanlastic mylonites).

OU82944/N100102M: low grade quartzose paramylonite (overprinting [presumably sil grade gneissic] Greenland Group/granite leucosome)

Hand: Deformed qtz dominant, silvery musc to 1 mm. Texture suggests low grade mylonite with anastomosing foliation and semi-brittle overprint. Rock cleaves on anastomosing foliation. py, ?chl?, ?ttn?

Thin: qtz (50%), microcrystalline birefr aggregate composed of recrystallised musc with relict coarser flakes enclosed (25%), fsp (20%), musc (5%), opaques

Textures are complex and heterogeneous. Grains fractured. Dominantly subgrain rotation recrystallization (SR) textures (sweeping undulose ext, deformation lamellae, subgrains, relatively straight grain boundaries, similar grain/subgrain sizes). Some small grains. Thin qtz ribbons with sutured margins common, also coarse (4-5 mm) highly def porphyroclasts of perthitic fsp, internally fractured and recrystallised along cross-cutting fracture system. Rare veins of calcite and patches of vermicular chl. Patches of recrystallized pl. Abundant qtz def bands cross cutting fabric.

Musc grains enclosed within microcrystalline recrystallized aggregate. Musc grains are 0.5 x 0.5 mm or 0.5 x 2 mm.
OU82942/N102801A: boudinaged foliation-parallel Greenland Group hosted qtz-fsp leucosome

**Hand:** fsp+qtz leucosome with weakly mylonitic texture (igneous texture largely intact), visible silvery musc to 4 mm, chl, ?zrc?; leucosome hosted within bio > musc quartzopelitic metasediment

**Thin:** qtz (50%), sericitised/albitised pl (45%), musc (3%), chl (2%), ap (1%)

Qtz has deformation bands, sweeping undulose extinction, dentate grain boundaries (though poor interconnectedness of qtz grains). Qtz grains are coarse (3 x 7 mm) with some sub-grain development along grain margins and along internal shear planes.

Chl (altered bio) occurs only near leucosome margin. Rounded, fractured ap grains (up to 0.9 mm).

Single musc grains to > 4 mm, internally deformed by kink planes. Chl not spatially associated with musc.

OU82945/N091503B1: med grade quartzo-pelitic paramylonite (overprinting [presumably sil grade gneissic] Greenland Group)

**Hand:** well-foliated mylonite with throughgoing segregations of qtz and ?ep/chl?, visible silvery musc is rare and small (< 0.5 mm), faint lineation due to mm-scale folds, cross cutting veins of qtz with small offsets

**Thin:** qtz (55%), microcrystalline aggregate (35%), musc (10%), pl (< 1%), chl?, 1 gar?

Texture suggests med grade mylonite, foliation has small scale folds but is otherwise continuous. Grain size is smaller and less variable then 02M sample. Some kinks and brittle offsets.
Qtz textures are dominantly SR. Sweeping undulose extinction, def lamellae, long ribbon grains, very fine subgrains. Larger ribbons 3 mm wide. Most qtz grains ~0.3 x 0.2 mm, some much smaller.

Musc grains are often small (0.1 x 0.3 mm) but some to 1 mm long. Lozenge shapes within fabric, in places as mica fish. Elsewhere associated with microcrystalline aggregate, but larger grains are clean.

Occasional qtz veins cross cut fabric.
A5. Fault Gouge Friction Experiment Plots
Refer to Table 2.2 for experiment and sample details.
A6. Carbonate C and O Stable Isotope Data

Refer to Supplement 2 for procedures, petrology and results.

---

**Iso-trace Analysis Report**

Results for: Nic Barth  
Sample set: Carbonates - Repeat analysis  
Date: 10/18/2011  
Iso-trace Job Number: 11606

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The following control material(s) were used to determine precision and accuracy:

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<td>Measured Values</td>
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A7. Paleontological Reports

A7.1 Amino Acid Racemization

Extent of valine and glutamic acid racemisation (total acid hydrolysate) in fossil molluscs from New Zealand

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<th>Species</th>
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<th>D/L glutamic acid</th>
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A7.2 Mollusca

D39/f0094, McKenzie Creek, South Westland

Bivalvia

Pecten novaezelandiae (Reeve, 1853), one complete, large, excellent LV, still with colour internally.
Talochlamys gemmulata (Reeve, 1853), several incomplete valves, some still brightly coloured.
Mytilidae, ?Perna canaliculus (Gmelin, 1791), 1 large umbonal fragment; identity uncertain.
Divalucina cuimingii (A. Adams & Angas, 1864), 3 good small valves.
?Paphies sp., 1 fragment of a smooth lower valve margin, identification uncertain (apparently from a “tuatua”, P. donacina (Spengler)).
Dosina zelandica Gray, 1835, 2 large fragments from a single valve.

Gastropoda

Maoricolpus roseus (Quoy & Gaimard, 1835), 1 good fairly complete shell.
Amalda (Baryspira) mucronata (G.B. Sowerby I, 1830), 3 fresh, broken to complete specimens.

Scaphopoda

Fissidentalium zelandicum (G.B. Sowerby II, 1860), 2 pieces from one moderately large shell.
Antalis nana (Hutton, 1873), 1 complete small specimen.

Brachiopoda (identified by Daphne Lee)

Terebratella sanguinea (Sowerby), one laterally crushed shell.

NCB Notes: From photos (seen but not collected): teredinid sp. (ship worms), Tana zelandica (gastropod), Austrofuscus glans (gastropod), Evechinus chloroticus (urchin)

Age: This fauna looks very young, just from the preservation, and the colour pattern remaining. But unfortunately, there is nothing about the macrofauna itself that is age-diagnostic, except that Pecten did not appear in the fossil record in NZ until c. MIS 31 (see Beu 2006, JRSNZ 36: 194 and following pp), and not at Wanganui until MIS 19. This specimen from Lake McKerrow is an unusual form with high but convex-crested ribs, resembling the “toi” (Fleming, 1957) form at Castlecliff, Wanganui, where it is limited to glacial MIS 14 (but that is not age-diagnostic, it merely means this was glacial, i.e., cold compared with Wanganui, I think).

Ecology: This is a shallow-water fauna, deposited in no more than ~10 m of water, probably in a quiet embayment (i.e., the earlier extension of Lake McKerrow). Pecten, Divalucina, Dosina, gastropods and scaphopods lived on or in a soft substrate in the bay, but most other taxa (mussel, Talochlamys and brachiopod) lived attached to hard substrates, so have been washed into this environment from a nearby hard substrate. This suggests the possibility of an exposure of hard rock on the Alpine Fault trace. The tuatua (if that’s what it is) has been washed into the deposition site from a nearby ocean beach, suggesting it might have lived on the ocean side of a sand bar alongside Lake McKerrow.

D39/f0095, Wolf River, South Westland

Bivalvia

Leionucula strangei, 5 valves (smooth interior margin).
Neilo australis, several excellent moulds & a few large fragments of shell.
Talochlamys gemmulata, several, sent to Colin Murray-Wallace for dating (some moulds and few fragments & juveniles remain in collection). Atrina zelandica, 2 small apices. Pratulum pulchellum, abundant, the most common species present. Serratina charlottae, quite common, small to large. Leptonychia retaria, several valves, most small. ?Tnaxia sp., one good internal mould; weakly truncate end, central umbo, no hinge teeth. Other smooth, subcircular bivalve – Felaniella sp., few poor.

Gastropoda
Tanea zelandica, one fairly large, fragile, crushed [checked this again later – very thin-shelled, i.e., possibly Globisinum], but I can see no spiral sculpture, so I don’t know what it could be if not Tanea. Only fragments remain after digging it out. Austrofusus glans, one quite large shell, and counterpart.

Scaphopoda
Cadulus sp. (sensu lato), 3-4 poor.

Echinoidea
?Pseudechinus, rather than Evechinus; several poor moulds of fragments of exteriors.

Remarks: Nothing here is diagnostic of age. The ecology represented is the infauna (and one specimen of the epifaunal carnivorous gastropod Austrofusus) of a very fine-grained muddy sea floor in c.10-50 m of water in a quiet, sheltered environment. Almost all taxa are filter-feeding infaunal bivalves, but Talochlamys is epifaunal (byssally attached to hard objects, evidently a nearby rock face) and Tanea (Naticidae) is a semi-infaunal carnivorous gastropod that pre dates infaunal bivalves. The echinoid Pseudechinus occurs now on soft bottoms in a wide range of depths.

D39/f096, Madagascar Creek, South Westland, 9-12 m above basement

Bivalvia
Nuculanidae, not determined; one smooth, oval valve. Talochlamys gemmulata, large sample, nearly all sent to Colin Murray-Wallace for dating, but several remain here. Pratulum pulchellum, a few valves.

Gastropoda
Cantharidus sp., one fragment, spire; still brightly coloured [as in the following sample].

Echinoidea
Pseudechinus sp. one quite good half-specimen, only interior visible.

Remarks: Again, nothing useful for age. The ecology is much more mixed than above – an almost equal mixture of the infaunal bivalve Pratulum and epifaunal taxa such as Talochlamys and Cantharidus. Evidently the silt was deposited near a rock face.

D39/f097, Madagascar Creek, South Westland, 15-17 m above basement

Bivalvia
Modiolus areolatus, 1 large valve, now fragmentary; mould complete. Talochlamys gemmulata, several small valves, all sent to Colin Murray-Wallace for dating. Limatula sutera, 1 small complete valve.
**Gastropoda**

*Haliotis virginea*, 1 almost complete shell (a RARE fossil!).
*Cantharidus dilatatus*, 2 good, still brightly coloured; very fine sculpture is diagnostic; they retain calcareous algae on the spire apex.
*Cominella (Josephba) powelli*, 3 good – the first fossil record.

**Cirripedia**

Two large barnacles, incomplete, only interiors visible; but the wide shape and large size indicate these are almost certainly *Austromegabalanus decorus*, a species that lives attached to molluscan shells in shallow subtidal water, i.e., not on intertidal rock faces.

**Crustacea**

In my opinion the isopod is a present-day terrestrial “slater” that has become stuck on the surface of the outcrop by accident. At first sight, it’s almost like a chiton with 12 “valves”, but all chitons have 8.

**Remarks:** Again, nothing diagnostic for dating, although the bright coloration of the trochid and the fact that some living species have not been reported fossil previously suggests it is very young. The ecology is very different from the above two samples, as ALL taxa present are epifaunal, and live attached to rocks and shells in about 5-10 m of water (with the possible exception of *Limatula suteri*). The silt evidently was deposited at the foot of a steep rock face, and its epifauna fell into the deposition site. A fiord-like site is suggested.

Alan Beu
a.beu@gns.cri.nz
4 April 2012
Modified slightly 17 April 2012
A7.3 Foraminifera

FORAMINIFERA IN UPLIFTED MARINE SEDIMENT, NEAR BIG BAY, SOUTH WESTLAND

Bruce W Hayward and Margaret S. Morley
Unpublished Report BWH 142/12
April 2012

Summary

The foraminifera in four samples from 400-580 m elevation near Big Bay, northern Fiordland have been analysed:

D39/0094 (N110610D): McKenzie Ck is entirely composed of living species, but the absence of *Saidovina karrerianum* could suggest that it is older than 0.34 Ma. It accumulated in sheltered normal marine conditions at a depth of 5-80 m, possibly at the shallow end of this, probably inside a fiord with an open entrance to the ocean.

D39/0094 (N110613A): McKenzie Ck is sandy and the foraminiferal tests partly dissolved, thus only large robust shells are preserved. It has a similar but lower diversity fauna to the sample above and possibly accumulated in shallower water (5-30 m).

D39/0095 (N120302C): Wolf contains rare, well-preserved specimens of six extinct benthic species that indicate an age older than 0.8 Ma. It accumulated in sheltered normal marine salinity conditions at a depth of 50-400 m, possibly 100-200 m depth inside a fiord.

D39/0096 (N120303C): Madagascar contains rare well-preserved specimens of four extinct benthic species and two extinct planktic species, and one late evolving planktic species that indicate an age of 1.8-0.8 Ma, possibly 1.8-1.5 Ma. It accumulated in sheltered normal marine salinity conditions at a depth of c.50-200 m, possibly inside a fiord.

Introduction

In March-April 2012, Nicolas Barth sent four samples of marine sediment from near Big Bay, northern Fiordland for an assessment of the paleo-environment and paleo-water depth of accumulation.

The material sent was:

N110610D One slide of picked foraminifera and ostracods “from Alpine Fault uplifted marine sediment” collected from McKenzie Creek, near Big Bay; thought to be MIS 7 (amino acid racemization dating), ~400 m above MSL.

N110613A One bag of washed sand from Mckenzie.

N120302C One bag of washed shelly sediment from Wolf, near Big Bay, northern Fiordland, ~580 m above MSL.

N120303C One bag of washed sediment from Madagascar, near Big Bay, northern Fiordland, ~580 m above MSL.

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### Species list

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<th>Mckenzie N110613A</th>
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<th>Madagascar N120303C</th>
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<td></td>
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<tr>
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<td>1</td>
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<td></td>
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<tr>
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<td>2</td>
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<td>Bolivina alata</td>
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<tr>
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<td>2</td>
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<td>9</td>
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<td>5</td>
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<td>Notorotalia inornata</td>
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</table>
Notorotalia zealandica  9  15  6  10-400 m
Oridorsalis umbonatus  7  50-4500 m
Praeglobobulimina spinosecens  1  80-500 m
Pseudopatellinoides primus  2
Rosalina bradyi  5  8  0-10 m
Rosalina irregularis  1  0-1500 m
Sphaeroidina bulloides  1  50-5000 m
Trifarina angulosa  2  2  25-2000 m
Uvigerina peregrina  1  50-5000 m
Total quant benthic forams  55  29  77  127

Planktics
Globigerina bulloides  2  5  5
Globigerina falconensis  2
Globigerina quinqueloba  1
Globigerinita glutinata  1
Globorotalia crassacarina  1  2.3-0.45 Ma
Globorotalia crassa  1  2.4-0 Ma
Globorotalia inflata  1  6  6  4.1-0 Ma
Globorotalia punctulata  1  5.3-0.7 Ma
Globorotalia truncatulinoides  2  1.8-0 Ma
Neogloboquadrina incompta  2  1
Neogloboquadrina pachyderma  1
Orbulina universa  1

Ostracoda
Munseyella brevis  1
Swansonites aequa  1
Trachyleberis scabrocuneata?  1
Trachyleberis sp. worn  1

Micromollusca
Gumina minor  1

Paleoenvironmental interpretation

D39/f0094 (N110610D) McKenzie Ck

The fauna from this sample does not contain the characteristic species that live in sheltered intertidal or shallow normal marine harbours or brackish inlets, lagoons or estuaries (e.g., Ammonia, Haynesina, Elphidium excavatum, E. charlottense, E. advenum; Hayward et al., 1999) nor does it contain any species that are restricted to living at depths greater than 50 m depth (Hayward et al., 2010). The fauna is typical of one that would be found today living in sediment of a relatively exposed coast at depths of 20-80 m, or possibly into shallower depths of as little as 5 m if they were more sheltered in a fiord or enclosed subtidal bay, like Port Pegasus (Hayward et al., 1994) and not subject to major turnover and reworking by waves. It has some similarity to the Notorotalia-Elphidium novozelandicum association that inhabits 5-20 m depths around the fringes of sheltered normal marine Port Pegasus today (Hayward et al., 1994, 1999). Thus it could have accumulated in the shelter of a fiord or bay that was fully marine with a relatively deep-water (possibly silled) opening to the ocean.
On exposed coasts like the West Coast of the South Island, foraminiferal and ostracod tests are absent or at least highly abraded if they accumulate on an exposed beach or in depths above the usual wave base. Clearly this fauna does not represent this kind of environment.

I conclude that a wide range of inner-mid shelf depths is possible, but most likely at the shallow end of 5-80 m.

The fauna includes two well-preserved specimens of a foraminiferal species (*Pseudopatellinoides primus*) not previously recorded from New Zealand. The Recent fauna from New Zealand is now pretty well known, so it is less likely that it is living today, especially as the genus has not been recorded from sediment younger than Pliocene (though it cannot be ruled out as a possibility). The species *P. primus* was described from Miocene strata overseas, which suggests that the specimens could be reworked from Neogene strata, such as those that outcrop around the entrance to Big Bay. The specimens are well-preserved which does not favour this reworking hypothesis. If these specimens were reworked from such strata, then maybe more of the fauna is also. Apart from *P. primus*, all other species are known to be living around NZ today and some postdate the Late Miocene in their time ranges. The explanation for the presence of *P. primus* in this deposit remains unresolved but possibly it was an interglacial immigrant that disappeared again during colder times.

D39/r0094 (N110613A) McKenzie Ck

This fauna is sparse and partly dissolved, such that only larger thick-shelled tests are present. It is dominated by *Notorotalia – Elphidium novozealandicum*, ND *Cibicides dispers*. Its modern analogue would be the *Notorotalia-Elphidium novozealandicum* association that inhabits 5-20 m depths around the fringes of sheltered normal marine Port Pegasus today (Hayward et al., 1994, 1999). This is supported by the presence of large *Rosalina bradyi* which live attached to shallow water seaweeds and rocks.

D39/r0095 (N120302C) Wolf

This fauna is dominated by the benthic foraminifera *Evolvocassidulina orientalis, Astronion novozealandicum, Oridorsalis umbonatus* and *Nonionellina flemingi*. Faunas with this composition have not been recorded around NZ today, but studies in Fiords have been limited to six faunas documented by Kustanowich (1965). Common *E. orientalis* and *N. flemingi* occur at 30-500 m water depth today, whereas the other two common species can be common right down to abyssal depths. Some possible constraints on the shallow end of the possible depth range comes from *O. umbonatus* (>50 m), *Globobulimina turgida* (>70 m) and *Pragglobobulimina spinescens* (>80 m) (Hayward et al., 2010). The extinct group benthics have usually been studied in deep-sea cores and are regarded as living primarily at middle bathyal and greater depths (>600 m). Since the specimens are rare in this fauna maybe it is possible that they lived at shallower depths, perhaps as shallow as 100-200 m in darker fiord bottom environments. Planktics comprise 17% of the total foraminiferal fauna indicating accumulation beneath neritic water, most likely at depths less than 100 m unless their abundance was affected by unusual fiord-like conditions.

I conclude that this fauna accumulated at 50-400 m water depth, possibly 100-200 m in the bottom of a deep fiord.

D39/r0096 (N120303C) Madagascar

This fauna is dominated by *Anomalinoides spherica* (36% of benthics) with common *Nonionellina flemingi, Quinqueloculina, Rosalina bradyi, Cibicides dispers, Fissurina lucida* and *Notorotalia*. Faunas with this combination have not been recorded from modern situations around New Zealand. The fauna has more shallow elements than Wolf and possibly accumulated at shallower depths or had shallower specimens transported into it. *A. spherica* has a peak of maximum abundance between
20 and 80 m today (Hayward et al., 2010) which is consistent with most of the common species. *Rosalina bradyi* is most common attached to algae or rocks at shallow depths of 0-30 m (Hayward et al., 1999), but possibly might have dropped or been transported into the fauna. Some rarer species provide insights into potential upper depth limit of this fauna (from Hayward et al., 2010): *Bolivina alata* (>100 m), *Cassidella brady* (>200 m), *Globocassidulina minuta* (>50 m), *Sphaeroidina bulloides* (>50 m), and *Uvigerina peregrina* (>50 m). Once again there are several extinction group benthics that usually occur in deep-sea cores at middle bathyal or greater depth, but rare specimens could potentially have lived in the unusual darker waters in the bottom of a fiord maybe as shallow as 100-200 m.

The 13% planktic foraminifera indicates neritic water overhead and depths most likely less than 75 m, unless in an unusual fiord setting.

I conclude that there is some conflicting evidence but the most likely compromise is that the faunas accumulated in an unusual deep fiord setting, at a depth of 50-200 m.

**Age**

As the samples were all originally thought by Nic to be uplifted Middle-Late Pleistocene in age (MIS 9-7) it was not thought that the shallow water benthic foraminiferal faunas expected would contain any age diagnostic species. This was the case in the McKenzie Ck samples but surprisingly not the case in the two higher elevation samples which both contained age diagnostic benthic foraminifera belonging to the group of bathyal and abyssal benthic foraminifera that became extinct globally in the mid-Pleistocene climate revolution (1.2-0.55 Ma). The time ranges used here are the global extinction times of species, although in many instances these are also their highest occurrences in the SW Pacific as well (Hayward et al., in press). There was also age diagnostic planktic foraminifera in the Madagascar sample and their age ranges used here are those summarised in Cooper et al. (2004).

**D39/h0094 (N110610D and N119613A) McKenzie Ck**

These faunas contain no species that are not living today. The species with the youngest first occurrence is *Globorotalia inflata* (FO 4.1 Ma). If the inferred paleoenvironmental setting is correct, then a modern fauna in this setting would have common *Saidovina karrerianum* (Kustanowich, 1965; Hayward et al. 1994). This species is recorded as having first arrived in New Zealand at c.0.34 Ma, based on onland outcrops at Te Piki (MIS 7) and Wanganui (Powell’s Wc locality). The lack of this species in these samples could indicate that it is older than 0.34 Ma, but there is no positive faunal evidence of an older age to support this, but this is unlikely to be present in such a shallow environment.

**D39/h0095 (N120302C) Wolf**

The presence of single, rare, but well-preserved specimens of several members of the global MPT extinction group of bathyal-abyssal benthic foraminifera (Hayward et al. in press) suggests an age older than 0.8 Ma. The only lower limit on the age is that of *Gr. inflata* (FO 4.1 Ma). These Ext. Gp benthics with their global extinction levels are:

*Bolivina pliozea* >0.46 Ma

*Orthomorphina jedlitschkai* >0.63 Ma

*Stilostomella fistuca* >0.7 Ma

*Mucronina spatulata* >0.7 Ma

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It is considered highly unlikely that all these six species are reworked, because of their excellent preservation and lack of other likely reworked species. They are however at the very shallow limits of their bathymetric range, which might explain their rarity.

D39/f0096 (N120303C) Madagascar

Like the previous sample this one has a few extinct benthic foraminifera (Hayward et al., in press) that constrain the upper age of the sample. These are:

- **Bolivinita pliozea**: >0.46 Ma
- **Siphonodosaria lepidula**: >0.57 Ma
- **Strictocostella advena**: >0.8 Ma
- **Hauserella pliozea**: >1.5 Ma

The highest occurrence of latter NZ endemic species is not well constrained in land sections because it mostly lived at upper bathyal depths. Its youngest recorded occurrence is in a continuous upper bathyal sequence in ODP 1119, Canterbury Basin, where it is well-dated at 1.5 Ma (Wilson et al., 2005).

In this sample there are also some planktic foraminifera whose age ranges (Cooper, 2004; Scott et al., 1990) constrain the age of this fauna:

- **Globorotalia crassacarina**: 2.3-0.45 Ma
- **Globorotalia crassula**: 2.4-0 Ma
- **Globorotalia inflata**: 4.1-0 Ma
- **Globorotalia puncticulata**: 5.3-0.7 Ma
- **Globorotalia truncatulinoides**: 1.8-0 Ma

Thus I conclude that there is strong evidence for an Early Pleistocene age within the span 1.8-0.8 Ma and likely 1.8-1.5 Ma, Wn.

**Update 03 September 2012 (following discovery of E. Huxleyi in samples):**

I am more happy now that the forams must be reworked – have you got a potential deepwater marine sedimentary rock source for the reworking near your site? or has it been eroded away? So I am now happy with an interpretation that has the foram fauna mixed - partly near in-situ fauna and partly reworked deeper water extinct benthics and a few reworked planktics. This is what I was saying to Daphne Lee in Brisbane.
References


A7.4 Nannofossils

Examination of Calcareous Nannofossils from Uplifted Alpine Fault Samples

Denise K. Kulhanek (d.kulhanek@gns.cri.nz)
October 2012

Methods

Four samples from Alpine Fault uplifted marine sediments were prepared for calcareous nannofossil examination following standard smear-slide techniques (e.g., Bown and Young, 1998). A small amount of sediment was scraped onto a coverslip from a fresh surface of each sample using a razor blade. The sediment was mixed with a drop of distilled water and spread evenly over the coverslip, dried on a hotplate, affixed to a glass microscope slide using Norland Optical Adhesive 61, and cured under an ultraviolet light. Slides were examined at 1000x and 630x using an Olympus BX53 light microscope under cross-polarized and brightfield illumination. A minimum of two coverslip traverses (approximately 400 fields of view) were observed for each slide and either presence/absence data collected (McKerrow Lake sample) or a qualitative assessment of the total abundance of nannofossils and individual species estimated as follows:

- A = abundant (>10 specimens per field of view).
- C = common (1 to 10 specimens per field of view).
- Fr = frequent (1 specimen per 2 to 10 fields of view).
- F = few (1 specimen per 11 to 50 fields of view).
- R = rare (1 specimen per >50 fields of view).
- B = barren (no specimens observed in 400 or more fields of view).
- * = reworked (reworked occurrence of that taxon).

In addition, a qualitative assessment of calcareous nannofossil preservation was assessed as follows:

- G = good (little or no evidence of dissolution and/or overgrowth, primary morphological characteristics only slightly altered, and specimens are identifiable to the species level).
- M = moderate (specimens exhibit some etching or recrystallization, primary morphological characteristics somewhat altered, but most specimens are identifiable to species level).
- P = poor (specimens severely etched or overgrown, primary morphological characteristics significantly altered, specimens may be largely fragmented and many specimens are unidentifiable at the species and/or generic level).

Results are correlated to the biostratigraphic zonation scheme of Martini (1971). Taxonomic concepts for species are those given in Perch-Nielsen (1985) and Bown (1998).

Results and Discussion

Of the four samples examined for calcareous nannofossils, one sample (D39/f0097) was barren. The other three (D39/f0096, D39/f0095, D39/f0094) contain moderately preserved assemblages ranging in overall abundance from common to abundant. The assemblages contain typical Pleistocene taxa including various species of *Gephyrocapsa* (*G. caribbeanica*, *G. oceanica*, *G. muellerae*, *G. ericsonii*, and *G. parallela*), *Calcidiscus leptoporus*, *Helicosphaera carteri*, *Umbilicosphaera sibogae*, *Rhabdosphaera clavigera*, and *Syracosphaera pulchra*. The presence of *Emiliania huxleyi* constrains the age to nannofossil Zone NN21, less the 290 ka (Gradstein et al., 2012).
Sample D39/f0095 also contains a number of taxa interpreted as reworked, based on the presence of *E. huxleyi*. Most of the reworked specimens have ranges into the Pleistocene, but go extinct before *E. huxleyi* evolves at 290 ka, including *Pseudoemiliania lacunosa* (last appearance datum [LAD] at 440 ka), large (>5.5 µm) *Gephyrocapsa* spp. (range: 1.62–1.24 Ma), and *Helicosphaera sellii* (LAD at 1.26 Ma). In addition, larger specimens of *Reticulofenestra baqi* are known from the Neogene. The absence of these taxa in the other samples examined, together with the presence of *E. huxleyi* within the assemblage, strongly supports the reworked interpretation.

The absence of nannofossils in sample D39/f0097 is interesting, as it is located stratigraphically ~5 m higher in the section than sample D39/f0096. The absence could be due to diagenesis; however, this is unlikely considering the geologically young age of the material and the presence of nannofossils in other nearby samples presumed to be of similar age. More likely, either the depositional environment excluded nannoplankton (e.g., terrestrial, marginal marine, or very nearshore) or a high energy environment may have winnowed fines, thus preferentially removing nannofossils during deposition of the sediment.

**References**


## Nannofossil Assemblages

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<td>B</td>
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| *Gephyrocapsa caribbeanica (<3 μm)* | C | R | | *
| *Gephyrocapsa caribbeanica (>6 μm)* | | | | *
| Gephyrocapsa ericsonii | X | R | | |
| Gephyrocapsa muellerae | X | Fr | Fr | |
| Gephyrocapsa oceanica | | A | Fr | R |
| *Gephyrocapsa oceanica (>4 μm)* | | | | |
| Gephyrocapsa parallela | X | R | | |
| Gephyrocapsa spp. (<3 μm) | C | C | | |
| Helicosphaera carteri | X | F | | |
| Helicosphaera sellii | * | | | *
| *Pseudoemiliania lacunosa* | * | | | *
| *Reticulofenestra haqii (3-5 μm)* | * | | | *
| *Reticulofenestra haqii (5-8 μm)* | | | | *
| *Reticulofenestra minutula* | F | R | | |
| Rhabdosphaera clavigera | X | R | | |
| Scyphosphaera spp. | X | | | |
| Syracosphaera pulchra | X | R | | |
| Umbellosphaera tenuis | F | | | |
| Umbilicosphaera cf. rotula | R | | | |
| *Umbilicosphaera sibogae* | X | R | | |
| Ascidian spicules | X | R | | |
| Calcispheres | X | F | | |

Notes: Abundance: A = abundant, C = common, Fr = frequent, F = few, R = rare, B = barren, X = present, * = reworked; Preservation: M = moderate
A7.5 Pollen

POLLEN ANALYSIS OF FOUR ISOLATED QUATERNARY SAMPLES FROM SHEET D39 –BIG BAY AREA

D.C. Mildenhall

GNS Science
Report DCM 505/12
13th June 2012.

Introduction

On two separate occasions in early 2012 I received a total of four samples from Nicolas Barth, who is studying the Alpine Fault in the area, to date and determine depositional environment. These were processed several months later in late May 2012. The samples were analysed in early June 2012. A full pollen list is attached.

The four samples were from sheet D39 and given the fossil record numbers D39/f094-097. Details of the localities and sediments sampled are on the fossil record forms.

Palynology

Although the four samples did not come from the same locality they are clearly very similar in age and environmental interpretation.

All samples are dominated by Cyathea spores and smooth monolette spores probably from a number of different ferns that produce such spores. The samples are probably water sorted with only f094 containing reasonably well-preserved spores and pollen. Counts for samples 096 and 097 are rather subjective with many mechanically broken and chemically etched, unidentifiable pollen and spore fragments.

Environment: D39/f094 is the only sample that indicates the local presence of forest trees; all other samples have Asteraceae (including a variety of daisy-type pollen), Poaceae (grasses) and other pollen types from shrubs and herbs dominating. All the other samples do contain tree pollen but in such low numbers that it is unlikely that extensive forest was anywhere near the sampled sites. The percentages of tree pollen present in the samples are 44% (094), 17% (095), 26% (096), and 17% (097), excluding spores in the percentages.

The samples are derived from a coastal acid swamp setting deposited into a marginally marine or lagoonal setting with a marine influence. The samples suggest cool to cold temperate conditions and high rainfall. Arcritarchs, presumed to be marine in origin, and/or marine dinoflagellate cysts were found in all sample bar 094. Charcoal commonly occurs in sample 097 which had the characteristics of a beach setting.

Age: There is little evidence of age but Haloragacidites harrisii (the fossil pollen of the she oak Casuarina) is present in all samples and in some numbers in sample 097. This would suggest an age of youngest Castlecliffian (c. 0.5-1.0 Ma). No other extinct taxa were located in a subsequent search. The pollen type appears to be in situ (I would expect other extinct pollen types if recycled – unusual for just one extinct pollen type to be selectively recycled) and is too frequent to be explained by pollen drift from Australia such as often occurs in Holocene West Coast sediments.
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D39/f0
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304
A8. New Zealand Fossil Record Electronic Database (FRED) Forms

Location
Field Number McKenzie Creek N110612F & N110613A
Original Grid Reference 1211775 5076078 (New Zealand Transverse Mercator)
Converted Grid Reference (D39) 2121863 5637797 (New Zealand Map Grid)
Converted Dec. Lat/Long 44.36623°S 168.12786°E (NZGD49)
Map Year Method GPS - Field
Accuracy 10.0 m
Locality Actively eroding slip gully on true left of the east branch of McKenzie Creek. Most samples collected within the bottom 5m of the marine unit.
Country New Zealand
Coordinate Comments
Locality Comments A secondary outcrop is located upstream of the east branch of McKenzie Creek also on the true left (about 175m due south).

Collection Information
Collector(s) Barth, Nicolas C.
Collection Date June 13, 2011
Fossils in Place Yes
Sent To (microflora) Mildenhall, D.C./GNS; pollen
(macrofauna) Murray-Wallace, C./Wollongong University NSW; Talochlamys fragments sent for amino acid racemization dating
(macrofauna) A. G. Beu/GNS; molluscs
(microfauna) Hayward, B.W.; forams
(microfauna) Kulhanek, D./GNS; nannos
Not Collected Photos taken: teredinid sp. (ship worms), Tanea zelandica (gastropod), Austrofuscus glans (gastropod), Evechiimus chloroticus (urchin)
Significance/Comments Quaternary marine sediment at 400m elevation adjacent to the Alpine Fault. Dating (in progress) will constrain a good long term uplift rate. Excellent preservation of wood (?podocarp bark?) and shells (fine ornamentation, mother-of-pearl and in some cases original colour is preserved) though shells are often fractured.

Stratigraphy
Stratigraphic Name
Inferred Stage
Known Stage
Samples Nearby
Sample Relationships
Strat. Relationships
Column/Map
Dip/Strike 20°NW/046° (Facing: Normal)
Stratigraphy Comments Marine unit is about 20m thick with well-preserved wood and shells abundant on horizons but also distributed throughout. Dropstones are also present. Underlain by >2m of locally derived brecciated ?Greenland Group? (most likely a scree/slip deposit). Overlain by >50m of moraine contiguous with the widespread glacial surface in the Jamestown Saddle area.

Sedimentary Features
Grain Size fine sand (pri), very fine sand (sec) (Comparator not used)
Bedding Thickness 5 - 60 cm

Printed on Tue Aug 21 16:08:52 NZST 2012 by Nicolas Barth from FRED, the computer database for the NZ Fossil Record File (FRF). D39/f0094
FRF is a nationally significant database administered by GSNZ and GNS Science Page 1 of 2
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<td>Colour</td>
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New Zealand Fossil Record File

Paleontological Record

**Field Number**

McKenzie Creek N110612F & N110613A

**Paleontology**

**Identifier(s):** Mildenhall, D.C.

**Identification Date:** May 2012

**Stage:**

**Stage Comments:**

**Lab Number:** 26554

**Collection Comments:**

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</tr>
<tr>
<td>Legarostrobos</td>
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<td></td>
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<tr>
<td>Malvaceae</td>
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</tr>
<tr>
<td>Melicytus</td>
<td>3</td>
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</tr>
<tr>
<td>Metrosideros</td>
<td>15</td>
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<tr>
<td>Myrsine</td>
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<td></td>
</tr>
<tr>
<td>Nothofagus menziesii</td>
<td>25</td>
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</tr>
<tr>
<td>Phyllocladus</td>
<td>3</td>
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<tr>
<td>Poaceae</td>
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<td>Podocarpus</td>
<td>4</td>
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<tr>
<td>Pseudopanax</td>
<td>3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tricolporates indet.</td>
<td>9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Weinmannia</td>
<td>3</td>
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</tr>
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cf. Nothofagus fusca
Location
Field Number: Wolf River N120302C
Original Grid Reference: 1195035 5063606 (New Zealand Transverse Mercator)
Converted Grid Reference: (D39) 2105144 5625349 (New Zealand Map Grid)
Converted Dec. Lat/Long: 44.46892°S 167.90887°E (NZGD49)
Map Year: 2012
Method: GPS - Field
Accuracy: 8.0 m
Locality: Outcrop of blue-grey silty sand in 3m high cliff in bush about 15m beyond head of the active Wolf River headscarp/slip. Samples collected about 3-4m stratigraphically above Greenland Group outcrop.
Country: New Zealand
Coordinate Comments
Locality Comments: Blue-grey silty sand is calcareous at depths >20cm from the surface. Other portions of the marine unit here (coarser sands) appear non-fossiliferous.

Collection Information
Collector(s): Turnbull, I.M.; Barth, Nicolas C.
Collection Date: March 2, 2012
Fossils in Place: Yes
Sent To: (microflora) Mildenhall, D.C./GNS: pollen
(macrofauna) A. G. Beu/GNS: molluscs
(macrofauna) Murray-Wallace, C./Wollongong University NSW: Talochlamys fragments sent for amino acid racemization dating
(microfauna) Kulhanek, D./GNS: nannos
(microfauna) Hayward, B.W.: forams
Not Collected
Significance/Comments: Quaternary marine sediment at 580m elevation adjacent to the Alpine Fault. Dating (in progress) will constrain a good long term uplift rate. Excellent preservation of shells (fine ornamentation and mother-of-pearl) though often fractured.

Stratigraphy
Stratigraphic Name
Inferred Stage
Known Stage
Samples Nearby: D39/f0039
Sample Relationships
Strat. Relationships: 3 m - 4 m above top GREENLAND GROUP
Column/Map
Dip/Strike: 5° SE/045° (Facing: Normal)
Stratigraphy Comments: Marine unit is about 18m thick with well-preserved shells distributed throughout. Dropstones are also present. Coarser sand portions appear non-fossiliferous, but display good bedding. Underlain by Greenland Group. Overlain by >5m of moraine presumably contiguous with the widespread Wolf River Tabellands glacial surface.

Sedimentary Features
Grain Size: very fine sand (pri), silt (sec) (Comparator not used)
Bedding Thickness: non-bedded
<table>
<thead>
<tr>
<th>Bedding Features</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Weathering</td>
<td>none or slight</td>
</tr>
<tr>
<td>Hardness</td>
<td>moderately soft</td>
</tr>
<tr>
<td>Carbonate</td>
<td>calcareous</td>
</tr>
<tr>
<td>Colour</td>
<td>medium blue-grey (Wet)</td>
</tr>
<tr>
<td>Additional Features</td>
<td>shelly; micaceous</td>
</tr>
<tr>
<td>Inferred Environment</td>
<td>Marine</td>
</tr>
<tr>
<td>Nature of Rock Unit</td>
<td></td>
</tr>
</tbody>
</table>

**Correspondence**
NEW ZEALAND FOSSIL RECORD FILE
Locality Record

Location
Field Number: Madagascar Creek N120303C
Original Grid Reference: 1193741 5063318 (New Zealand Transverse Mercator)
Converted Grid Reference: (D39) 2103852 5625052 (New Zealand Map Grid)
Converted Dec. Lat/Long: 44.47078°S 167.69244°E (NZGD49)
Map Year: 2012
Method: GPS - Field
Accuracy: 6.0 m
Locality: Great section of marine silts and sands exposed in slip area at the head of the next major drainage south of the Wolf River (referred to as Madagascar Creek). Fossils from 9 to 12m above basal contact with Greenland Group outcrop.
Country: New Zealand

Collection Information
Collector(s): Turnbull, I.M.; Barth, Nicolas C.
Collection Date: March 3, 2012
Fossils in Place: Yes
Sent To: (microflora) Mildenhall, D.C./GNS; pollen
(macrofauna) A. G. Beu/GNS: molluscs
(macrofauna) Murray-Wallace, C./Wollongong University NSW: Talochlamys fragments sent for amino acid racemization dating
(microfauna) Hayward, B.W.: forams
(microfauna) Kuhaneck, D./GNS: nannos

Not Collected:
Significance/Comments: Quaternary marine sediment at 530m elevation adjacent to the Alpine Fault. Dating (in progress) will constrain a good long term uplift rate. Excellent preservation of shells (fine ornamentation and mother-of-pearl) though often fractured.

Stratigraphy
Stratigraphic Name:
Inferred Stage:
Known Stage:
Samples Nearby: D39/f0097
Sample Relationships: 3 m - 6 m below D39/f0097
Strat. Relationships: 9 m - 12 m above top GREENLAND GROUP
Column/Map:
Dip/Strike: 18°SW/160° (Facing: Normal)
Stratigraphy Comments: Marine unit is about 25-30m thick with well-preserved shells distributed throughout (though most abundant and diverse in the middle of the section). Large dropstones are also present and increase in abundance upwards. Underlain by Greenland Group. Overlain by 35m of moraine contiguous with the widespread Wolf River Tablelands glacial surface.

Sedimentary Features
Grain Size: very fine sand (p1i) (Comparator not used)
Bedding Thickness: 5 - 50 cm
Bedding Features:

Printed on Tue Aug 21 08:14:00 NZST 2012 by Nicolas Barth from FRED, the computer database for the NZ Fossil Record File (FRF). D39/f0096
FRF is a nationally significant database administered by GNS and GNS Science
<table>
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<th>Property</th>
<th>Description</th>
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<tbody>
<tr>
<td>Weathering</td>
<td>none or slight</td>
</tr>
<tr>
<td>Hardness</td>
<td>moderately soft</td>
</tr>
<tr>
<td>Carbonate</td>
<td>calcareous</td>
</tr>
<tr>
<td>Colour</td>
<td>medium grey (Dry)</td>
</tr>
<tr>
<td>Additional Features</td>
<td>shelly; micaceous</td>
</tr>
<tr>
<td>Inferred Environment</td>
<td>Marine</td>
</tr>
</tbody>
</table>

**Correspondence**
### Location
- **Field Number**: Madagascar Creek N120303D
- **Original Grid Reference**: 1193756 5063305 (New Zealand Transverse Mercator)
- **Converted Grid Reference**: (D39) 2103867 5625049 (New Zealand Map Grid)
- **Converted Dec. Lat/Long**: 44.47091°S 167.89262°E (NZGD49)
- **Map Year**: 6.0 m
- **Locality**: Great section of marine silts and sands exposed in slip area at the head of the next major drainage south of the Wolf River (referred to as Madagascar Creek). Fossils from 15 to 17m above basal contact with Greenland Group outcrop (near large 1.5m wide drumline)
- **Country**: New Zealand

### Collection Information
- **Collector(s)**: Turnbull, I.M.; Barth, Nicolas C.
- **Collection Date**: March 3, 2012
- **Fossils in Place**: Yes
- **Sent To**:
  - (microflora) Mildenhall, D.C./GNS: pollen
  - (macrofauna) A. G. Beu/GNS: molluscs
  - (macrofauna) Murray-Wallace, C./Wollongong University NSW: Talochlamys fragments sent for amino acid racemization dating
  - (microfauna) Hayward, B.W.: forams
  - (microfauna) Kulhanek, D./GNS: nannos

### Stratigraphy
- **Stratigraphic Name**: D39/f0096
- **Inferred Stage**: Greenland Group
- **Known Stage**: Greenland Group
- **Samples Nearby**: 3 m - 5 m above D39/f0096
- **Strat. Relationships**: 15 m - 17 m above top GREENLAND GROUP
- **Column/Map**: (Facing: Normal)

### Stratigraphy Comments
- Marine unit is about 25-30m thick with well-preserved shells distributed throughout. Dropstones are also present and increase in abundance upsection. Underlain by Greenland Group. Overlain by 35m of moraine contiguous with the widespread Wolf River Tablelands glacial surface.

### Sedimentary Features
- **Grain Size**: very fine sand (pri) (Comparator not used)
- **Bedding Thickness**: 5 - 60 cm
<table>
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<tr>
<th>Bedding Features</th>
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<tbody>
<tr>
<td>Weathering</td>
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<tr>
<td>Hardness</td>
<td>calcareous</td>
</tr>
<tr>
<td>Carbonate</td>
<td>medium grey (Dry)</td>
</tr>
<tr>
<td>Colour</td>
<td>shelly, micaceous</td>
</tr>
<tr>
<td>Additional Features</td>
<td>Marine</td>
</tr>
<tr>
<td>Inferred Environment</td>
<td></td>
</tr>
<tr>
<td>Nature of Rock Unit</td>
<td></td>
</tr>
<tr>
<td>Correspondence</td>
<td></td>
</tr>
</tbody>
</table>
**NEW ZEALAND FOSSIL RECORD FILE**

**Locality Record**

**Locality Comments**

**Collection Information**

Collector(s) | Turnbull, I.M.; Nathan, S.; Beu, A.G.
Collection Date | March 1978
Fossils in Place | Almost
Not Collected | (macrofauna) Beu, A.G./NZGS
Significance/Comments | COLLECTION OF MACROFOSSILS FROM LARGE BOULDERS IN LOWER REACHES OF WHISKEY CREEK. FOSSILIFEROUS BAND PROBABLY OCCURS ABOUT 250 M UPSTREAM BUT NOT SEEN IN SITU.

**Stratigraphy**

Inferred Stage | Terangian - Aranuian (0.43 - 0.0 Ma)

**Sedimentary Features**

Grain Size | mud (pri), very fine sand (sec)
Bedding Thickness | none or slight
Hardness | moderately soft
Carbonate | calcareous
Colour | light grey-blue (Wet)
Additional Features | micaceous; rock fragments
Inferred Environment | Marine
Nature of Rock Unit | Marine

**Correspondence**

**Masterfile**: Southern Sth Island  
**Approved**: Ben Morrison October 13, 2008  
**Comments**: Approved after backlog editing

**Location**

- **Field Number**: BNT-5
- **Original Grid Reference**: (E38) 2144800 5678800 (New Zealand Map Grid)
- **Converted Dec. Lat/Long**: 44.00967°S 169.44358°E (NZGD49)
- **Map Year**: 1975
- **Accuracy**
- **Locality**: LARGE BOULDERS IN LOWER REACHES OF WHISKEY CREEK COAST BETWEEN TEER CREEK AND GIANTSHIP. NORTH COAST OF CASCADE POINT, SOUTH WESTLAND.
- **Country**: New Zealand

**Stratigraphic Name**

- **Known Stage**
- **Samples Nearby**
- **Sample Relationships**
- **Strat. Relationships**
- **Column/Map**
- **Dip/Strike**
- **Stratigraphy Comments**

**Outcrop**

**FOSSIL RECORD NUMBER**

E38/f0015


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315
**Field Number**

**Paleontology**

**Identifier(s)**
Beu, A.G.

**Identification Date**
March 30, 1978

**Stage**
Terangan - Aranuian (0.43 - 0.0 Ma)

**Stage Comments**
NO EXTINCT TAXA; ALL VERY FRESHLY PRESERVED, SOME RETAIN COLOUR; DEFINITELY POST-CASTELELFIAN; ?LAST GLACIATION??.

**Lab Number**
GNS GS 12260

**Collection Comments**
A SHALLOW, IN-SHORE FAUNA, NOT FAR OFF A STEEP ROCKY COAST - A FIORD SITUATION?

### Brachiopoda

**Taxonomic Name**
- Magasella sanguinea

### Bivalvia

**Taxonomic Name**
- Aulacomya maoriana
- Chlamys gemmulata suteri
- Chlamys zelandiae
- Leptomya retia
- Modiolus areolatus
- Nemocardium pulchellum
- Panopea smilliae
- Paphinus largillierti
- Perna canaliculus
- Scalpmacra scapellum
- Tawera 'aff. bollonsi
- Thracia vitrea

### Gastropoda

**Taxonomic Name**
- Buccinulum sp.
- Callistoma sp.
- Cirsotrema zelebori
- Cominella (Eucominio) nassoides
- Miorelenchus
- Trichosiris octocarinatus
- Uberema vitrea
- Xyrene ambiguus
- Xyrene aucklandicus
- Xyrene convexus
- Xyrene mortenseni caudatinus
- Zegalerus tenuis

### Annelida

**Taxonomic Name**
- gen. indet.
<table>
<thead>
<tr>
<th>Taxonomic Name</th>
<th>Spec Count</th>
<th>Spec Coord</th>
<th>Comments</th>
</tr>
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<tbody>
<tr>
<td>Echinoidea</td>
<td>Pseudochinus sp.</td>
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<td></td>
</tr>
<tr>
<td>Gen. indet.</td>
<td></td>
<td></td>
<td>2 species of crabs</td>
</tr>
</tbody>
</table>
NEW ZEALAND FOSSIL RECORD FILE
Locality Record

FOSSIL RECORD NUMBER
E38/f0037
www.fred.org.nz/locality/E38/f0037
Outcrop

Location
Field Number
Original Grid Reference
Converted Dec. Lat/Long
Map Year
Method
Accuracy
Locality
Country
Coordinate Comments
Locality Comments

Collection Information
Collector(s)
Collection Date
Fossils in Place
Sent To
Not Collected
Significance/Comments
Regional extent of Teer Formation. Regional uplift, offset on the Alpine fault.

Stratigraphy
Stratigraphic Name
Inferred Stage
Known Stage
Samples Nearby
Sample Relationships
Strat. Relationships
Column/Map
Dip/Strike
Stratigraphy Comments

Sedimentary Features
Grain Size
Bedding Thickness
Bedding Features
Weathering
Hardness
Carbonate
Colour
Additional Features
Inferred Environment
Nature of Rock Unit

Correspondence

Printed on Tue Oct 11 09:23:42 NZDT 2011 by Nicolas Barth from FRED, the computer database for the NZ Fossil Record File (FRF).

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Page 1 of 1
**Field Number**  SP2

**Paleontology**

**Identifier(s)**  Beu, A.G.

**Identification Date**  July 6, 1993

**Stage**  Waipipian (3.6 - 3.0 Ma)

**Stage Comments**  Post Waipipian, nothing extinct (if Teer Formation, presumably last interglacial or early last glacial, from stratigraphy).

**Lab Number**  GNS GS 14962

**Collection Comments**  A fine sediment fauna that could have lived in any depths from c. 10 metres on down. But N.B. very low diversity.

<table>
<thead>
<tr>
<th>Taxonomic Name</th>
<th>Spec Count</th>
<th>Spec Coord</th>
<th>Comments</th>
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<tbody>
<tr>
<td>Bivalvia</td>
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</tr>
<tr>
<td><em>Limatula ?suteni</em></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><em>Nemocardium (Pratulum) pulchellum</em></td>
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<td></td>
</tr>
<tr>
<td><em>Zenalia acinaces</em></td>
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</tr>
<tr>
<td>Scaphopoda</td>
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<td></td>
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<tr>
<td><em>Cadulus sp.</em></td>
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</tr>
<tr>
<td>Echinodermata</td>
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<td></td>
</tr>
<tr>
<td><em>gen. indet.</em></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Comments**  Numerous small -> large fragments, from more than one valve.

Abundant, most crushed.

One juvenile valve, c. 20 mm long.

Small, smooth, polished shells, one large kept, 2 frags. in fine fraction.

A few small, flat-section spines in fine fraction (? Echinocardium, or similar).