Surface breakthrough of a basement fault by repeated seismic slip episodes: The Ostler Fault, South Island, New Zealand

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[1] The Ostler Fault is one of the major active reverse faults in the piedmont of the Southern Alps, SE of the Alpine Fault. We present a new geological and morphotectonic map of the southern Ostler Fault, integrated with two seismic reflection profiles across the active central segments of the fault. Segmented, subparallel scarps define a N-S belt (~40 km long and 2–3 km wide) of pure reverse faults, which upthrow and back-tilt a panel of Plio-Pleistocene terrestrial units (2.4–1.0 Ma) plus the overlying glacial outwash (<200 ka). Uplift gradients, the chronology of newly faulted markers, and tectonically controlled diversion of paleodrainages, all indicate progressive S to N breakthrough of the surface trace of the Ostler Fault in the last 2.4 Ma. The new seismic data define a main fault segment dipping 50°–60°W to depths of ~1.5 km, with a vertical throw of 800 m, and a shortening of ~30%. The fault geometry and kinematics and the subsurface data favor the interpretation that the Ostler Fault propagated updip across the Plio-Quaternary terrestrial sequence as the emerging, high-angle splay of an inherited Late Cretaceous–Paleocene normal fault, that underwent repeated cycles of compressional reactivation in the last 2.4 Ma.


1. Introduction

[2] The active tectonics of the South Island of New Zealand is governed by convergence between the Australian and Pacific plates, with the largest components of shortening being accommodated at the Puysegur subduction margin, along the transpressive, right-lateral interplate boundary of the Alpine Fault, and within the array of right-lateral fault splays in the Marlborough region (Figure 1a).

[3] Together, the largest faults account for between 50% and 90% of the total plate motion [Norris and Cooper, 2001; Sutherland et al., 2006], with the remaining fraction apparently distributed into a wider fault network off the plate boundary [Walcott, 1978]. The seismic potential of reverse faults away from the plate boundary is well demonstrated by the location and magnitude [Anderson and Webb, 1994; Eberhart-Phillips and Reyes, 1997] of the two largest earthquakes of the last century in the South Island (Murchison, 1929, M, 7.8; and Inangahua, 1968, M, 7.4, Figure 1a). Sets of N-S to NE-SW long-lived faults in both the Australian and Pacific crust date back to the Late Cretaceous–Paleocene extensional phases [Bishop and Buchanan, 1995], and a suite of structural and paleoseismic data [Bishop and Beanland, 1991; Jackson et al., 1996; Barrell and Cox, 2003; Ghisetti and Sibson, 2006] demonstrate that many have indeed undergone compressional reactivation in the present stress field, under an axis of maximum horizontal stress oriented 295° ± 16° [Balfour et al., 2005]. Recognition of the seismic potential of a large set of basement faults that often lie buried beneath cover sequences poses serious problems for seismic hazard assessment, especially when the recurrence interval between successive earthquakes is longer than the historical record. Localization of inherited basement faults requires a broad range of subsurface data, and in many cases evidence is collected after the occurrence of a major earthquake on blind, or previously unmapped faults [Stein and King, 1984; Namson and Davis, 1988; Stein and Ekstrom, 1992].

[4] In the piedmont of the Southern Alps, the N-S oriented Ostler Fault (Figure 1b) is one of the best structures for deciphering the progressive surface breakthrough of a reverse fault active during sedimentation of a Plio-Quaternary terrestrial sequence, incorporating glacial outwash, that overlies the strongly deformed basement of the Pacific Plate east of the Alpine Fault. A number of studies have documented the control exerted by the Ostler Fault on Late Pleistocene terrestrial sedimentation and landscape evolution in the Mackenzie Basin [Davis et al., 2005; Amos and Burbank, 2007; Amos et al., 2007], and the repeated surface rupture of the fault during M > 6 earthquakes in the Holocene [Read, 1984; Van Dissen et al., 1993; Davis et al., 2005]. In contrast, the long-term geological evolution of the Ostler Fault (especially the relationship between newly propagating faults that have repeatedly ruptured the surface and reactivated, deep-seated and potentially seismogenic basement faults) has not been extensively investigated so far.

[5] In this paper we present a set of new geological, structural and morphotectonic data of the southern Ostler Fault system between Omarama and Twizel (Figures 2 and 3), combined with the interpretation of two seismic reflection profiles across the most active, central strands of the fault (at Lake Ohau Road and Willowbank Saddle). Together,
surface and subsurface data make it possible: (1) define a set of subparallel reverse faults that penetrate the basement with dips of $\sim 50^\circ - 60^\circ W$ down to depths of $\sim 1.5 \, \text{km}$; (2) constrain a maximum vertical throw of 800 m along the major, easternmost strand of the Ostler Fault, and a total shortening of $\sim 30\%$ of the Plio-Quaternary terrestrial sediments across a belt of localized deformation $\sim 2 \, \text{km}$ wide, and (3) reconstruct the progressive deformation and the south-to-north propagation of the surface breakthrough of the Ostler Fault over the last 2.4 Ma.

[6] The offset of Plio-Quaternary sediments over a localized surface fault trace $>40 \, \text{km}$ long requires repeated cycles of activity, associated with seismic rupture of basement faults at crustal depths $>5 \, \text{km}$, and indicates that the Ostler Fault plays a significant role in the present tectonic regime in accommodating shortening away from the plate boundary. The N-S to NNE-SSW strike of the Ostler Fault system lies nearly perpendicular to maximum horizontal stress trajectories and to the contemporary direction of shortening inferred for the northern and central South Island [Beavan and Haines, 2001; Balfour et al., 2005]. This suggests that movement across the Ostler Fault involves close-to-pure reverse dip-slip, consistent with the lack of evidence for significant strike-slip across the structure. In turn, this raises the issue as to whether the Ostler Fault is a new, optimally oriented reverse fault that started to propagate during the late Pliocene and is still growing, or is rather the emergent splay of an inherited basement fault that has been selectively reactivated over the last 2.4 Ma.

[7] These two alternatives carry different implications for the assessment of the seismic potential of the Ostler Fault, in terms of the geometry of the fault plane relative to the contemporary stress field, its depth of penetration, and extent of an eventual rupture during a large earthquake, and also the recurrence interval between large earthquakes.

[8] Our data cannot solve the geometry of the seismo-genic master fault, but (1) the moderate to high-angle dip of the fault down to 1.5 km depth, (2) the fabric of N-S basement faults inherited from earlier extensional episodes, and (3) the presence of an inverted sedimentary basin in the fault hanging wall inferred from the gravimetric minimum in the Mackenzie Basin [Kleffmann, 1999] lead us to speculate that the Ostler Fault is the surface breakthrough of an inherited normal fault that has been compressionally inverted during the Quaternary. As such, the Ostler Fault is only one of the many basement faults adjacent to the Alpine Fault plate boundary capable of seismic reactivation in the present stress field, for which the recent rupture history is revealed by its propagation through, and progressive disruption of, a well-preserved sequence of Plio-Quaternary terrestrial sediments and fluvioglacial markers.

2. Geological Setting

[9] The Mackenzie Basin (Figure 2) is the northernmost and most extensive ($\sim 1700 \, \text{km}^2$) of a series of intramontane Neogene-Quaternary terrestrial basins in the eastern foothills of the Southern Alps. In the north Otago region, coarse clastic deposits fill narrow, elongated depressions bounded by NE-SW range-front reverse faults, active during the Late Quaternary [Youngson et al., 1994; Jackson et al., 1996]. The Mackenzie Basin is crossed by the Ostler Fault, an $\sim 50\,-\text{km}$-long, active reverse fault that extends S-N from Omarama to Twizel, and, farther north, along the eastern margin of the Ben Ohau Range, terminating $\sim 25 \, \text{km}$ SE of...
Figure 2. (a) Geological map of the Ostler Fault between Omarama and Twizel, simplified from the original map surveyed at the scale of 1:25,000. See Figure 1 for regional location. Note the intrabasinal upthrusting of the Late Pliocene–Pleistocene (2.4–1.0 Ma) terrestrial deposits of the Kurow Group (L1, L2, and L3) in the hanging wall of the fault (see also Figure 4). Dashed traces along cross sections 1 and 2 mark the traces of the Lake Ohau Road and Willowbank Saddle seismic lines. Corresponding geological cross sections are in Figure 8. (b) Composite stratigraphic column, providing also the key for the geological map. Note the change of vertical scale between the Rakaia terrane basement, the terrestrial deposits of the Kurow Group, and the glacial and fluvio-glacial deposits 186 to 16 ka old. Also note that the chronostratigraphic order of superposition of the river terraces is in inverse order to their relative elevation above the basin floor.
the Alpine Fault (Figure 1b). In the area between Omarama and Twizel, a new geological map at scale 1:25,000 was surveyed by F. C. Ghisetti as the base document for two seismic lines shot by A. R. Gorman and a crew of students in the early months of 2006, and processed in the geophysical laboratory of the Geology Department at the University of Otago. A simplified and reduced version of the geological map is presented in Figure 2.

Figure 3. Structural and morphotectonic map of the Ostler Fault between Omarama and Twizel. Areal extent is the same as in Figure 2. Topographic contours and hydrography are redrawn from the topographic map of New Zealand at a scale of 1:50,000 (Land Information New Zealand, www.linz.govt.nz). Note the set of subparallel faults and folds that define the principal fault trace (OF1), and the west stepping of faults west of Mt. Ostler along the trace of OF2. Uplift of the Plio-Pleistocene Kurow Group and of the glacial outwash in the fault hanging wall is marked by the alignment of hills rising ~400 m above the basin floor, the development of enclosed basins, and the abandoned drainages in suspended wind gaps. See text and Figure 5 for further details.
Figure 4. Geometry of the sequence exposed along the Ostler Fault in the mapped area. (a) Subvertical layers of sandstones and argillites of the Permian–Late Triassic Torlesse turbidites on the western slopes of Benmore Range. South of Lake Ruataniwha. (b) Bedded Torlesse greywackes exposed on top of Table Hill and interpreted as a sliver of upthrusted basement in the hanging wall of the Ostler Fault. (c) W-dipping sequence of fluvio-lacustrine silty claystones, mudstones, quartzose sandstones, and quartz conglomerate. Lower Kurow Group (L1 in Figure 2a) exposed at Clay Cliffs, W of Omarama. (d) Lenses of quartz gravels (L2 in Figure 2a) at the gradational transition to the upper gravels of the Kurow Group, N of Quailburn. (e) Clast supported, bedded gravels in the uppermost Kurow Group (L3 in Figure 2a) at Clay Cliffs, W of Omarama. (f) Sequence exposed in the hanging wall of the Ostler Fault at Mt. Ostler, NW of Twizel. Note the growth geometry of the W-tilted well-bedded L3 unit of the Kurow Group (base of the cliff) and the gentle westward dip of the erosional unconformity underlying a package of coarse gravels attributed to Terrace T1. (g) Surface folding of the Terrace T3 in the hanging wall of OF2, S of Trig Z (the fault scarp is a few meters east of the left margin of the photo). Folding is accompanied by sets of minor N-S faults with both reverse and normal components, developed in the hinge zone of the anticline.
Terrane), interbedded with brown mudstones, argilites and green-red shales [Gair, 1967]. A metamorphic transition from the Torlesse turbidites to more deformed foliated metagreywackes and low-grade schists has been traced in the Diadem Range, at the western margin of the Mackenzie Basin [Forsyth, 2001]. At the southern end of the Ostler Fault (Figure 2a), low-grade schists are faulted against the Late Pliocene–Pleistocene terrestrial sequence, and foliated schists, upthrust in the hanging wall of the Ostler Fault are exposed at Table Hill (Figure 4b), and on the eastern flank of the Ben Ohau Range (Trig Z, Figure 2a). The basin contact of the terrestrial sediments with the Torlesse basemant is nowhere exposed. Given the proximity to the Tertiary marine basins in South Canterbury (Figure 2) it has been speculated that the deepest part of the Mackenzie Basin may host Late Cretaceous–Tertiary marine sequences buried beneath the thick blanket of terrestrial Neogene and Quaternary cover [Gair, 1967].

[11] The terrestrial sequence preserved in outcrop [Speight, 1940; Spörli and Lillie, 1974] comprises a lower unit with ~200 m of exposed thickness (Figure 4c) starting with green-white silty clayslstones, mudstones and quartz sandstones interbedded with quartz conglomerates (L1), probably deposited in a braided fluvial to lacustrine environment. These deposits pass gradationally upward to 5- to 50-m-thick lenses of interbedded coarse quartz gravels and silts (L2) (Figure 4d), overlain by clast-supported rusty-brown bedded gravels (>300 m thick), with pebbles and cobbles of greywackes and semischists (L3) (Figure 4e). The age of this terrestrial sequence (Kurow Group) is poorly constrained [Mildenhall, 2001] to Late Pliocene–Pleistocene (~2.4 to 1.0 Ma). Provenance indicators are rare, but at a few sites imbrication in the gravels is consistent with deposition from prograding alluvial fans and braided rivers descending from the eastern slopes of the Southern Alps.

[12] In the hanging wall panel of the Ostler Fault the whole sequence of the Kurow Group is uplifted and backtilted 30°–60°W. The tilted beds are truncated by an erosion surface (Figure 4f), dipping 0°–10°W. Above this surface rest matrix-supported gravels (Figure 4f), with dominant pebbles and cobble-size greywacke clasts, deposited at different elevations in a sequence of downcutting aggradational episodes. Stratigraphic ties (Figure 2b) are provided by: (1) correlation with the glacial deposits of the Mackenzie Basin [Gair, 1967]; (2) relative topographic elevation of the terraced surfaces, and, (3) superposition of nested deposits. Deposits attributed to the Waimean glaciation (186–128 ka) are exposed at the highest elevation above the basin floor [Read, 1984; Barrell and Cox, 2003] where they interfinger (e.g., Table Hill and Willowbank Saddle) with the gravel belongings to the oldest and highest terraces (T1 and T2 in Figures 2a and 2b). Three younger events (Figures 2a and 2b) are identified by distinct festoons of moraines and fluvo-glacial outwash, with: (1) terrace T3 correlated with the Balmoral 2 (71–59 ka) glacial deposits; (2) terrace T4 correlated with the Mt. John (24–18 ka) deposits and, (3) the youngest set of nested terraces (T5 a, b, c) correlated with the progressive downcutting of the Tekapo (18–16 ka) glacial outwash along the Ohau River [Matzels, 1989; Suggate, 1990; Barrell and Cox, 2003] at the outlet of the glacial Lake Ohau. The total thickness of the glacial and Holocene alluvial cover varies locally from tens of meters to >200 m [see also Long et al., 2003].

3. Ostler Fault Zone

[13] Between the Ahuriri River and Twizel the surface trace of the Ostler Fault zone consists of a W-stepping array, ~40 km long and 2–3 km wide, of subparallel, discontinuous faults, trending N-S to NE-SSW (Figure 3). Different local names have been given to these faults in the literature, but here all the fault splays at the front of the uplifted terrestrial sequence are interpreted as belonging to the principal Ostler Fault 1 (OF1). The northern, W-stepping fault west of Mt. Ostler is named Ostler Fault 2 (OF2), and continues farther north along the west shore of Lake Pukaki (Figure 1b), where metagreywackes and schists are thrust above a terrestrial sequence similar to the one exposed in the Mackenzie Basin [Smith et al., 1996].

[14] The faults have average dips of ~50°–60°W at the surface [see also Amos et al., 2007] and they offset the Plio-Pleistocene terrestrial units of the Kurow Group and the whole sequence of glacial deposits and correlative outwash, 186–128 to 18–16 ka old. Uplift of units in the hanging wall panel (Figures 2a and 3), ground deformation monitored by geodetic leveling [Blick et al., 1989], and paleoseismic analyses [Van Dissen et al., 1993] all indicate reverse movement, with shortening nearly perpendicular to the fault system. No component of strike-slip displacement is apparent.

[15] The fault splays of OF1 are marked by closely spaced topographic scarps down-stepping to the east, up to 100 m high in the older units and 1–20 m high in terraces T3 and T4. The whole terrestrial sequence L1, L2, L3 (Figure 2) is everywhere back-tilted 30°–60°W in the hanging wall of OF1 (see also Figures 4c, 4d, 4e, and 4f). Only minor evidence of hanging wall folding is preserved (e.g., synclines and anticlines with axial traces sub-parallel to OF1 at Clay Cliffs and at the north end of Table Hill; see Figures 2 and 3). At Mt. Ostler, the beds of the L3 gravels show an internal geometry of fanning strata (Figure 4f), consistent with depositional growth over the rotating hanging wall panel of OF1 [see also Amos et al., 2007]. Also, the surfaces of terraces T1 to T4 at Clearburn, north of Lake Ruataniwha, and west of Mt. Ostler are buckled by asymmetric anticlines with meter-scale normal faults in their crests and steep eastern forelimbs truncated by the OF1 and OF2 scarps (Figures 2, 3, and 4g). There are no offset Plio-Pleistocene markers to quantify directly the total throw of OF1. However, the decrease in topographic relief and exposed vertical thickness of the uplifted terrestrial sequence in the fault hanging wall are consistent with the south-to-north decrease in fault displacement, associated with the westward step to OF2. Slip rates estimated from the offset of deposits 120–22 ka old [Read and Blick, 1991] range between 0.4 and 1.5 mm/a.
On the basis of variations in slip rate, previous authors [Read, 1984; Van Dissen et al., 1993; Davis et al., 2005] have argued for segmentation of the fault system, with a southern segment nearly 15 km long (Clay Cliffs-Willowbank Saddle), a central segment nearly 11 km long (Willowbank Saddle-Clearburn) and a northern segment nearly 12 km long (Table Hill-Mt. Ostler) (Figure 3). However, there is no clear surface evidence for discontinuities separating the inferred segments. The Ostler Fault system abruptly terminates southward [cf. Read, 1984] against NNW-SSE faults that juxtapose L3 against the schists of the Diadem Range (Figures 2 and 3). These faults trend subparallel to the Otematata fault system (Figure 1b), an inherited Late Cretaceous normal fault that underwent compressional inversion in the Neogene [Bishop, 1976].

### 4. Geomorphic Expression of the Ostler Fault

The erosive landforms and the hydrographic network of the Mackenzie Basin are strongly controlled by the glacial and postglacial cycles of the last 200 ka, but significant geomorphic anomalies occur close to the Ostler Fault, testifying to disruption imposed by recent faulting. The relatively subdued topography of the intramontane Mackenzie Basin is interrupted by the linear ridge of the uplifted and W-tilted Plio-Pleistocene sequence L1-L2-L3,

![Figure 5](image-url)

**Figure 5.** Paleodrainage carved on top of Terrace T4 (correlative with the 24–18 ka Mt. John outwash) in the hanging wall of the central segment of the Ostler Fault, in the wind gap of Clearburn. See Figure 3 for regional location. The currently dry channels of the paleodrainage have been mapped on a georeferenced set of air photos at a scale of 1:25,000. The morphology of the channels shows the reversal of earlier E-flowing drainages, captured by younger, W-flowing drainages that are tributaries of the Ohau River. Note the location of an active spring in the footwall of OF1, and the deflection of the secondary divide in the folded hanging wall (refer to Figure 3 for the axial trace of the folds). See text for further details.
Table 1. Seismic Acquisition Parameters

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that pierces the alluvial cover in the hanging wall of the Ostler Fault system. From Clay Cliffs to Clearburn (Figure 3), the morphology is dominated by four aligned hills that define a sinuous intrabasinal divide, culminating at 893 m asl. This secondary divide separates narrow river valleys that flow to the east across the fault scarp (tributaries of the Ahuriri River), from a poorly evolved westward drainage captured by subcircular enclosed basins that are trapped between the W-tilted hanging wall panel and the eastern slopes of the Diamond Range (Figure 3).

[18] At Willowbank Saddle, remnants of the 186–128 ka old outwash mark the site of a now abandoned and suspended transverse paleovalley (wind gap), formerly carved by a drainage flowing eastward across the fault. The thick outwash deposits are faulted in the hanging wall of OF1, uplifted nearly 100 m relative to the present river valley, and warped into an anticlinal bulge with the west limb back-tilted ~5°W.

[19] At Clearburn, the terrace T4 correlative with the Mt. John glacial deposits (24–18 ka; see also Figure 2) marks the position of the paleodrainage that once flowed transverse to OF1 [cf. Read, 1984]. Braided paleochannels on the top surface of T4 were presumably carved during the last stages of late glacial to interglacial retreat [Maizels, 1989]. Detailed mapping of this channel network from air photos at 1:25,000 scale (Figure 5) shows a late, W-flowing drainage superposed onto a previous E-flowing drainage, with the earlier channels truncated and captured by the cloverleaf of the Diamond Range (Figure 3), the morphology is dominated by four aligned hills that define a sinuous intrabasinal divide, culminating at 893 m asl. This secondary divide separates narrow river valleys that flow to the east across the fault scarp (tributaries of the Ahuriri River), from a poorly evolved westward drainage captured by subcircular enclosed basins that are trapped between the W-tilted hanging wall panel and the eastern slopes of the Diamond Range (Figure 3).

5. Seismic Reflection Profiles and Fault Geometry at Depth

5.1. Acquisition and Processing

[21] In January and February 2006, a crew from the University of Otago acquired two seismic reflection lines across the Ostler Fault (Willowbank Saddle and Lake Ohau Road seismic lines, see Figure 2a and Table 1). The source points for the survey were 150-g explosive charges buried to depths of 0.7–1.2 m in holes excavated by a power auger. Geophones for the 48-channel survey were spaced 10 m apart. The ~9.5-km-long northern line along Lake Ohau Road (section 1 in Figure 2a) had a shot spacing of 40 m (sixfold coverage), whereas the ~8.0-km-long southern line at Willowbank Saddle (along the trace of section 2 in Figure 2a) had a shot spacing of 60 m (fourfold coverage). Shot and geophone locations were surveyed using a differential Global Positioning System (GPS).

[22] Seismic processing made use of the University of Otago’s seismic processing facilities, consisting of Globe Claritas software [Ravens, 2001] running on a Linux platform. The data were first compiled and geographically positioned using GPS. A crooked line geometry was used to assign common-midpoint (CMP) bins. Noisy traces were systematically removed and a debias filter was applied. Difficulty in drilling to significant depths in the glacial outwash gravels led to shallow charges and undesirable blowouts from many of the shots. The resulting airwave was reduced in amplitude through a partial muting procedure.

[23] Significant near-surface variability and topography necessitated the careful application of refraction and residual static corrections. First breaks on all shot gathers were picked using an automatic pick utility that was manually overridden as needed. Traveltimes were then inverted to produce a well-constrained near-surface velocity model from which surface-consistent static corrections could be determined. At this stage in the processing, the refraction and residual static corrections were applied relative to a floating datum located on the ground surface; this has less effect on the near-hyperbolic shape of reflections and should produce a better velocity analysis and stack.

[24] Because coverage was low, velocity analyses were performed primarily using constant velocity stacks. Near surface velocities were set by the results of refraction static analysis. The velocity model and resulting stacked data (e.g., Lake Ohau Road line, Figure 6) are well constrained where reflective energy is coherent (to about a maximum of ~1 s just east of the surface trace of OF1). The velocity model provides a first-order indication of greater thicknesses of shallow and young low-velocity sediments east
of the surface trace of OF1. The effect of upthrusted older higher-velocity units in the central part of the profile is also seen. For convenience, the stacking velocity was set to 5000 m/s at the bottom of the seismic data (4 s).

[25] Post-stack processing commenced with the application of post-stack statics (determined earlier as part of the refraction static analysis) to convert from a floating datum to a horizontal datum (600 m above sea level for the Lake Ohau Road line). Following this, a Butterworth band-pass filter with corner frequencies of 15–30–120–150 Hz was applied. The data were then migrated with a finite difference time migration routine using an interval velocity model constructed from the stacking velocities using the Dix equation. F-X domain deconvolution was then applied to improve lateral coherency of reflections and to reduce random high-amplitude noise. Finally, the data were converted from time to depth using the same interval velocity model used for the migration.

5.2. Geological Interpretation

[26] The Lake Ohau Road seismic line has significantly better resolution than the Willowbank Saddle seismic line, and it is the only one presented here (Figure 7), though both seismic lines were processed, depth converted and interpreted (see Figure 8).

[27] In the Lake Ohau Road seismic line (Figure 7a) the only tie between seismic reflections and outcropping geological units is provided by the package of high-energy reflections, ≤300 m thick, dipping 30°W, imaged at ~200 m below the surface between CDP 550 and 600. From position, elevation and angle of dip this package corresponds to the Plio-Pleistocene silty claystones and mudstones (L1) that outcrop in the hanging wall of OF1, immediately south and north of the line (Figure 2a). The most prominent surface trace of OF1 sharply truncates these reflections between CDP 500 and 550 and can be traced with dips ~50°–60°W down to ~1 km below surface (Figure 7b).

[28] In the footwall of OF1 a package of layered, reflective sediments onlaps a nonreflective unit along a boundary that can be projected into outcrop as the nonconformity at the top of the Torlesse greywackes. The westward inflection of this boundary from ~400 m below surface at the eastern end of the line to ~1000 m below surface at the footwall cutoff of OF1 appears to be accommodated by a couple of W-dipping normal faults, and delineates a growth basin in the footwall of OF1, filled with a sequence ≤800 m thick. This sequence should include, from bottom to top, all units of the Kurow Group (L1, L2 and L3, Figure 2), plus the uppermost cover of glacial and alluvial deposits. According to this interpretation, the basement is upthrusted ~200 m below surface in the hanging wall of OF1, and the total, post L1 reverse throw of OF1 amounts to ~800 m (Figure 7b). The faulting of the whole Plio-Quaternary sequence is accomplished by the main strand of OF1 and by a shallow footwall shortcut (OFS1) that truncates the uppermost horizons and emerges at the surface at CDP 400. Another splay emerging at surface at CDP 600 (OFS2) is traced west of the main trace of OF1, with ~100 m of reverse throw in L1. The branching of fault splays from OF1 is consistent with the multiple scarps mapped at surface, and with the deformation of terrace T4 between CDP 500 and 600 (Figures 2 and 3).

[29] West of the main trace of OF1, the geological interpretation (Figure 7b) is based on the downdip projection of the base of L3 (striking 020° and dipping 30°W) that is in outcrop at an elevation of 540 m asl in the hills immediately south and north of the line (Figure 2a). With this geometry, the base of L3 should be intersected at ~300 m below
The topmost reflective unit is interpreted as L3, the poorly reflective unit below it as L2, and the lowest package of west-dipping high-energy reflections that overlie the nonreflective basement as L1. Between CDP 700 and 850, L3 is faulted and folded by sets of minor reverse faults (F3, F4, F5, F6) with maximum vertical throws \( \leq 100 \text{ m} \).

West of CDP 950, the lower part of the seismic section loses resolution, but the sharp truncation of the W-dipping reflections attributed to L1, L2 and L3 is likely caused by a reverse fault dipping 55\(^\circ\) W. This inferred fault has no obvious surface trace, but it can be projected along strike to merge with the trace of OF2 mapped farther north (Figures 2a and 3). The hanging wall of this westernmost fault has an unclear seismic definition, but the topmost subhorizontal reflections (400 m below surface) correlate with glacial and alluvial outwash strata that greatly hinder the penetration of seismic energy to deeper horizons. Below, faint reflections are mostly evident between CDP 1000 and 1300, and show a possible unconformity of subhorizontal beds (L1?) truncating W-dipping reflections down to 1 km below surface. However, the poor resolution of the line and the lack of surface ties make it impossible to provide a sound geological interpretation at this end of the line.

![Figure 7](image)

**Figure 7.** (a) Migrated and depth-converted Lake Ohau Road seismic line and (b) corresponding geological interpretation. CDP spacing is 5 m. L1, L2, and L3 are the terrestrial units of the Plio-Pleistocene Kurow Group. See text and Table 1 for a description of the seismic processing and of the constraints used to tie surface data to subsurface reflections. Note the inflection of the sedimentary basin in the footwall of OF1, and the set of high-angle, subparallel reverse faults that define a 2- to 3-km-wide belt of intense shortening between OF1 and OF2. West of OF2 there is no adequate constraint to interpret the faint, west-dipping reflections underneath the glacial outwash.

The geological interpretation of the subsurface geometry of the Ostler Fault, as constrained by the surface mapping and the seismic reflection profiles is given for two cross sections, one coincident with the Lake Ohau Road seismic line (section 1, Figure 8a) and the other incorporating the Willowbank Saddle seismic line (section 2, Figure 8c). Though surface and subsurface deformation is much better constrained along section 1, similar structural characters are also recognized in section 2, especially concerning:
1. Surface splaying of a subsidiary fault (OFS1) from the major strand of OF1, dipping 45\(^\circ\) W;
2. A reverse throw of \( \approx 800 \text{ m} \) on OF1, calibrated on L1; and
3. Draping of the sedimentary panel above blocks of the rigid basement upthrust in the hanging wall of OF1. In both sections 1 and 2, the westernmost fault is tentatively correlated with OF2, but there is no clear surface evidence for a continuous fault trace extending as far south as section 2.

The geometrical admissibility of the geological interpretation of the Lake Ohau Road seismic line (Figure 8a) has been tested by restoring the L1, L2 and L3 units above the greywacke basement (after removal of the overlying glacial and alluvial cover, and updip projection above the present erosion surface). Restoration has been performed with the software 2D Move (Midland Valley Exploration
assuming mechanisms of flexural slip of the clastic cover detached above the Torlesse greywacke. The restoration (Figure 8b) produces a geometrically consistent retrodeformed section, with a shortening of ~30% largely ascribed to OF1 and its splays OFS1 and OFS2. Note however that restoration of L3 to an inferred original bedding of \( C_{24} \) does not remove all of the deformation within L1 imposed by earlier faults (both normal faults and the reverse fault F3) and by folding localized along the propagation trajectory of OF1.

6. Discussion

[33] Between Omarama and Twizel the two major strands of the Ostler Fault system (OF1 and OF2) together with their splays define a set of subparallel reverse faults dipping 50°–60°W from the surface down to ~1.5 km depth, within a belt ~2–3 km wide (Figures 2, 3, and 7). The lack of any obvious strike-slip across the principal components of the reverse fault system trending N-S to NNE-SSW constrains regional \( \sigma_1 \) trajectories to be oriented roughly E-W to WNW-ESE. This is consistent with the inferred stress trajectories oriented at 295° ± 16° in the northern South Island [Balfour et al., 2005].

[34] Bending and faulting is recorded by the progressive tilting of the sedimentary panel above the rigid substratum, as indicated by: (1) the diminishing angle of dip from 30°–60°W of the 2.4–1 Ma old L1, L2 and L3 units to 10°W of the pre-Balmoral 186–128 ka old gravels (terrace T1, see Figures 2 and 4f) that rest unconformably above the older sequence, (2) the growth geometry of the fanning beds of L3 in the hanging wall of OF1 (Figure 4f), and (3) the westward increase in thickness and the eastward onlap of the L1–L3 sedimentary sequence above the basement in the growth basin located in the footwall of OF1 (Figure 7). This geometry can result from either (1) progressive limb rotation during synsedimentary growth of fault-propagation folding at the tip of the upward propagating OF1, coeval with sedimentation of the Kurow Group.
anticlines [see also Amos et al., 2007] or (2) compressional oversteepening of faulted panels above a common detachment horizon. In favor of the second alternative is the lack, both in the field and in the subsurface, of extensive folding in the Plio-Pleistocene terrestrial deposits. However, there is clear evidence of localized antclinal buckling of terraces T1 to T4 in the hanging wall of OF1 and OF2. The overall geometry of these anticlines, with the crestal uplift immediately adjacent to the fault scarp, a steep east limb, and a gently W-dipping back-\(\text{limb, is mimicked by the deformation measured by repeated levelings along the central part of OF1 [Blick et al., 1989], with rates of hanging wall uplift of 0.6–0.7 mm/a, decreasing to zero at a distance of 400–800 m west of the fault.}

[35] Thus it is conceivable that folding in the older sediments of the Kurow Group is poorly preserved as a consequence of the offset of early propagation folds by repeated breakthrough of the upward propagating fault, accompanied by erosion and/or collapse of the poorly consolidated and unconfined sediments in the steeply dipping east limbs.

[36] All surface and subsurface data (Figure 8) are consistent with persistent pulses of reverse reactivation of the Ostler Fault system during the whole interval of deposition of the Kurow Group (loosely constrained as 2.4 to 1.0 Ma old), with (see also Figure 8b): (1) initiation of folding and reverse faulting post deposition of L1 in a localized belt between OF2 and OF1; (2) repeated episodes of fault growth and surface breakthrough localized along OF1 and its splay during deposition of L3, resulting in the increased W-tilting of the hanging wall panel and in the localization of a growth basin in the deflected footwall filled by the topmost gravels of L3 (Figure 7); and (3) pulses of activity on both OF1 and OF2 after the deposition of the glacial gravels \(\sim 186–16\) ka old. During these stages reverse folding and faulting have been accompanied by localized normal faulting that accommodates stretching in the crests of the anticlines (Figure 4g), and the flexural bending in the footwall of OF1 (Figure 7).

[37] The interpretation of the subsurface data allows us to define a total reverse throw of \(\sim 800\) m along the central segment of OF1 (calibrated over the offset of L1 in Figure 8a), and the restoration of the deformed Plio-Pleistocene sequence (Figure 8b) gives a total horizontal shortening of \(\sim 30\%\) for the Plio-Pleistocene deposits faulted between OF1 and OF2. These data constrain an average slip rate of 0.3–1.0 mm/a for the central segment of OF1 over the whole succession of Plio-Pleistocene terrestrial deposits (2.4–1.0 Ma). These long-term slip rates provide only a minimum estimate, but they are in the range of those calculated from the displacement of the younger Balmoral, Mt. John and Tekapo surfaces (0.4–1.5 mm/a according to Blick et al. [1989] and 1.1–1.7 mm/a according to Amos et al. [2007]), especially considering the error margins inherent to these time-averaged slip rates.

[38] It is also notable that displacements (and, consequently slip rates) change along the Ostler Fault from south to north. The highest topographic relief and the largest exposed thickness of the terrestrial sequence uplifted in the hanging wall of OF1 occur along the southern strands of the fault, from Clay Cliffs to Willowbank Saddle, and progressively decrease northward, from Willowbank Saddle to Mt. Ostler (Figures 2a and 3). Accordingly, the age of newly faulted markers becomes younger from the central segments (where the fault displaces the 186–128 ka deposits) to the northern segments (where the fault displaces terraces T3, T4 and T5).

[39] In addition, geomorphic evidence of (1) abandoned drainages in the uplifted wind gaps at Willowbank Saddle (terrace T1) and Clearburn (terrace T4) and (2) reversal from E- to W-directed cross-fault flow on the surface of terrace T4 at Clearburn are consistent with progressive abandonment and shifting of the E-flowing pre-Balmoral (186–128 ka) and Mt. John (24–18 Ka) drainages transverse to the Ostler Fault (Figures 3 and 5), eventually captured by the Ohau River after the Tekapo (18–16 ka) glaciation. This evolution can be related to westward backtilting of the topographic slope (<\(10^\circ\)), caused by antclinal folding and surface upthrusting along progressively younger segments of OF1 from south to north. However, the fault-controlled topography has been repeatedly defeated over time by rivers that have carved new valleys across the active fault scarp, (e.g., Quailburn, Clearburn and Ohau Rivers, Figure 5).

[40] Thus surface and subsurface data (Figures 7 and 8) show: (1) decreasing amounts of shortening, tilting and offset from the older (~2.4–1 Ma) units of the Kurow Group to the younger glacial and fluvioglacial deposits 186–16 ka old, connected with breakthrough of OF1 across fault propagation folds (Figure 8b); (2) progressive S to N migration of the break of the Ostler Fault, since deposition of the pre-Balmoral outwash (186–128 ka); and (3) repeated formation of temporary morphological barriers caused by discrete episodes of folding and faulting during the aggradational episodes of terraces T1 (186–128 ka) to T4 (24–18 ka).

[41] All these elements indicate progressive south-to-north breakthrough of the Ostler Fault during repeated seismic slip episodes through the Quaternary.

[42] Surface bend folding above steep thrust faults buried at shallow depth [e.g., Stein and King, 1984], and coseismic lateral propagation of surface breaks during rupture of blind reverse faults are well documented, and have been related to stress transfer at the fault tips, accompanied by distributed secondary faulting [e.g., Lin and Stein, 2004]. Surface propagation of antclinal ridges exerts a strong control on the diversion of hydrographic networks [Burbank et al., 1996; Jackson et al., 1996; Delcaillau et al., 1998; Keller et al., 1999], with development of centrifugal drainages, wind gaps and bowl-shaped alluvial basins (see numerical models of Champel et al. [2002]) similar to those observed along the Ostler Fault (Figure 3).

[43] The observed relationships established between earthquake magnitude, rupture length, and surface displacement for thrust earthquakes [Wells and Coppersmith, 1994; Lettis et al., 1997] show that the probability of surface rupture increases at magnitudes \(M \geq 5.9\), and that the ratio of surface to subsurface rupture length is typically \(\sim 0.75\). Instrumentally monitored seismic activity in the Mackenzie
Figure 9. Schematic interpretation of the Ostler Fault system to a depth of 5 km, i.e., the shallowest likely depth for nucleation of M ≥ 6 earthquakes. West of OF2, the basin beneath the glacial outwash is inferred from a marked gravity low of ~25 mGal (Kleffmann and Stern, manuscript in preparation, 2007), but is not imaged by our subsurface data, apart from faint, W-dipping reflections in the Lake Ohau Road seismic line (Figure 7). OF1 and OF2 are depicted as the youngest surface splays of a steep normal fault, compressionally inverted in the Quaternary-to-present tectonic regime. This interpretation emphasizes: (1) the localized deformation in a 2- to 3-km-wide belt along the Ostler Fault; (2) the maintenance at depth of the high-angle dips of the faults; and (3) the position of the Ostler Fault system at the boundary between an eastern block of uplifted basement, and a western block that hosts an earlier, extensional basin (Tertiary marine?), with an apparent null displacement consequent on reversal of slip on the reactivated normal fault.
50° ± 5° was attributed, at least in part, to compressional inversion of former normal faults.

[45] Dip estimates obtained for the principal components of the Ostler Fault system (50°–60°) thus lie close to the upper bound observed for reverse fault ruptures approaching the expected angle for frictional lock-up, and are consistent with the faults having originated as normal faults that subsequently underwent reactivation during compressional inversion. In this interpretation (Figure 9) the reversal of slip on the inverted normal fault has resulted in an almost zero net displacement of the greywacke basement and of the former Late Cretaceous–Tertiary extensional basin (marine?) prior to sedimentation of the L1–L3 sequence and updip propagation of the youngest, shallow strands of OF1 and OF2.

[46] In agreement with other field examples and analogue experiments [cf. Buchanan and Buchanan, 1995] our reconstruction of the geological history of the Ostler Fault in the last 2.4 Ma shows that: (1) the propagation of new faults is not favored when the inherited fabric provides a large set of structures suitably oriented for reactivation; (2) reactivation of inherited basement faults is highly selective; and (3) once reactivation of an individual fault strands occurs, it persists over long-lasting and repeated cycles of activity, leading to updip propagation and eventually to surface break.

[47] The basement in the Mackenzie Basin and adjacent regions [Cox and Barrell, 2007] is disrupted by sets of steep faults with similar orientations to the Ostler Fault (Figure 1b). All of them are potential candidates for accommodating components of shortening in the hanging wall of the Alpine Fault. However, their history of activity and/or compressional inversion during the Pli–Quaternary cannot easily be demonstrated because of jagged mountainous terrain coupled with the lack of preserved sedimentary cover.

7. Conclusions

[48] The Ostler Fault is one of the most active subsidiary structures proximal to the central strand of the Alpine Fault, with an orientation nearly orthogonal to inferred trajectories of maximum horizontal stress. Surface faulting and folding of the Late Pliocene–Quaternary terrestrial sequences (Figures 3, 5, 8, and 9) provide a record of repeated pulses of almost pure reverse fault movement along the Ostler Fault.

[49] The evaluated maximum throw of ~800 m along the central strand of the Ostler Fault, and the 30% of total shortening in the Late Pliocene–Pleistocene terrestrial units (2.4–1 Ma old) were largely acquired before deposition of the 186–128 ka outwash gravels, but folding and faulting continued repeatedly over the last 120 ka. Over this time interval, fault activity has successfully competed with the erosive processes governed by successive glacial episodes and with the aggradation in the foothills of the rapidly uplifting Southern Alps, resulting in a significant morphostructural imprint and in temporary diversion of the drainage system (Figure 5).

[50] Decreasing amounts of shortening, tilting and offset from the older to the younger deposits, and south-to-north propagation of the surface trace of Ostler Fault appear to mark the progressive breakthrough of shallow faults spaying from a buried fault. Full exposure of the fault traces mapped at the surface (Figure 3) presumably occurred during a large number of rupture events (M ≥ 6) spanning the whole Late Pliocene–Quaternary.

[51] We lack adequate subsurface mapping and seismicity to define the geometry of the seismogenic fault in the basement of the Mackenzie Basin, but it seems probable that inherited basement structures with N-S orientation play a major role absorbing shortening off the plate boundary, and we favor the interpretation (Figure 9) that the faults of the Ostler Fault system are the youngest, updip segments of a steep inherited (Late Cretaceous–Paleocene) normal fault, inverted in compression as a consequence of E-W shortening during the Late Pliocene–Quaternary.

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