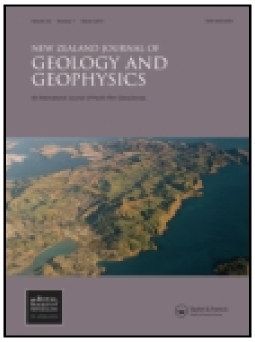

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The tectonic and structural setting of the 4 September 2010 Darfield (Canterbury) earthquake sequence, New Zealand

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Plate boundary deformation creates a south-easterly advancing, repetitive structural pattern in Canterbury dominated by the propagation of northeast-striking thrust assemblages. This pattern is regularly segmented by east-striking faults inherited from reactivated Cretaceous normal faults. The more evolved and deeply exposed structures in the foothills of north Canterbury provide insights into the tectonic processes of the blind structures now emerging from under the northern and eastern Canterbury Plains, where thrust and strike-slip fault activity are closely linked. The east-striking faults separate relative motion between thrust segments and accommodate oblique transpressive shear. Early stages of thrust emergence are dominated by anticlinal growth and blind, or partially buried, thrusts and backthrusts. The east-striking transecting faults therefore record timing of coseismic episodes of uplift and shortening with variable horizontal to vertical ratios and displacement rates on the hidden adjacent thrusts. The Greendale and blind Port Hills faults, with their associated aftershock patterns, are compatible with this style.

Keywords: Ashley Fault; blind thrusts; Boby Stream Fault; Canterbury; fault propagation folds; Greendale Fault; Springbank Fault; strike-slip; structure; tectonics

Introduction

The surface rupture of the east-striking Greendale Fault on 4 September 2010, with its textbook mesoscopic-scale dextral strike-slip shear zone, properly focused attention on recording the detail. Both this and the subsequent 22 February 2011 Christchurch event, on the yet-to-emerge Port Hills fault, showed up anomalies that were inconsistent with simple strike-slip faults. Firstly, the surface rupture length of 29.5 km for the Greendale Fault appeared to define a fault plane too short for the high-energy release and up to 5.3 m displacement for the M_W 7.1 initiating earthquake (Quigley et al. 2012). A fault plane length of approximately half that and displacement of 2.5 m has been estimated for the latter M_W 6.2 event (Beavan et al. 2011), but is part of an equally long linear aftershock zone. Large single-event offsets relative to surface rupture length on other Canterbury faults have been noted in the palaeoseismic record (Campbell et al. 1994). A second anomaly is that thrust focal mechanisms are associated with both the M_W 7.1 and M_W 6.2 triggering events beginning each cycle of aftershocks dominated by strike-slip focal mechanisms (Gledhill et al. 2011; Kaiser et al. 2012). Of particular note is that the Charing Cross epicentre is c. 8 km north of the Greendale Fault rupture, and the offshore New Brighton M_L 5.8 and 6.0 and related thrust aftershocks is up to 10 km north of the Port Hills on 23 December 2011 (GeoNet 2011).

Transverse belts of aftershocks extending for many kilometres across the ends of the east-striking strike-slip faults were recognised as involving synchronous reverse faulting on several interconnected structures (Gledhill et al. 2011). As preliminary data became available, the consistency of aspects of this newly identified composite structure, with the characteristic structural style of known Canterbury faulting and tectonics, was evident within a few days of the 4 September 2010 Darfield earthquake sequence.

The differential interferometric synthetic aperture radar (DInSAR) and GPS-derived ground deformation pattern surrounding both the Greendale and Port Hills faults (Beavan et al. 2010, 2011) reflects the net strain from all active sources and extends inhomogeneously far into the wall rocks of the fault planes. Both ground deformation patterns show a similar broad, gently south-plunging, anticlinal uplift of the south walls approximately parallel to the length of the faults plus secondary anticlines coinciding with linear aftershock clusters, notably near Charing Cross and at the western end of the Greendale Fault (Beavan et al. 2010, 2011).

The link between the strike-slip faults and blind-thrust-generated anticlines is unsurprising as it is very consistent with the patterns well established by structural and geomorphic mapping over the northern Canterbury Plains and surrounding foothills. Because structures are progressively evolving and propagating to the south and east with time,

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both thrust and strike-slip faults become more emergent and deeply exposed further back into the foothills. Topographic and drainage anomalies are indicators of similar structures beneath the Quaternary outwash gravels of the Canterbury Plains, now confirmed by recently acquired seismic reflection data (Jongens et al. 1999; Estrada 2003; Finnemore 2004; Dodson 2009; Dorn et al. 2010).

Crustal shortening on northeast-striking thrust-fault blocks dominates the regional structural style and these interact with secondary dextral-transpressive east-striking faults. Since Nicol (1993a) first described the transpressive inversion of the Birch Fault, a Cretaceous normal fault, it has been recognised that other similarly oriented faults were probably reactivated inherited faults, a view now supported by seismic reflection data (Ghisetti & Sibson 2012; Jongens et al. 2012). The emergence of these faults as active surface traces is observed to be closely associated with the location of thrust faults and their associated anticlines.

In this paper we outline the known tectonics and structures of North Canterbury and explore the influence of dominant crustal shortening on the behaviour of the coeval strike-slip faults, equivalent to the Darfield earthquake sequence. This outline is built upon at least two decades of research at the University of Canterbury (UoC) including numerous student theses, much of which is unpublished.

The northeast-striking thrust system

The landwards projection of the Chatham Rise is deforming in response to the eastwards expansion of the plate boundary deformation zone, forming a distinctive structural assemblage. Evidence of a ductile mid-crustal detachment at about 12 km beneath the northern Canterbury Plains (Reyners & Cowan 1993), thinning to 10 km (Bannister et al. 2006) west of the Harper River, may separate westwards convergence of the lower crust from the brittle upper crust. Based on an analysis of a temporary array of seismographs, Cowan (1992) identified a southern transition to a more coherently integrated crust under the plains from near the Rakaia River. The response to detachment is the production of a series of upper plate internally shortened thrust blocks, each fronted by a southeast-facing fault zone propagating from the basal decollement (e.g. Bannister et al. 2006). Both synthetic shears and antithetic back thrusts further shorten and disrupt each block, forming mountain ranges bound by inwardly dipping faults (Cowan 1992; Chamberlain et al. 1996). The family of faults associated with each thrust block and associated frontal thrust is colour coded into groups in Fig. 1, as are other categories of faults according to their characteristics. West of Mottram Peaks, this pattern is lost through total overprinting by the north-northwest grain of faults associated with the main divide ranges. Tracing the thrust fault groups out to the Canterbury Plains margin (Groups 1, 2 and 3 of Fig. 1), the frontal thrusts are the west-dipping element of the Harper Fault zone (Chamberlain et al. 1996), the Torlesse and Big Ben range-front faults (Cowan 1992) and the Hororata Fault (Jongens et al. 1999). The spacing between blocks is very similar at approximately 25–30 km. Of similar dimensions is the step to the eastern end of the Greendale Fault and again to the eastern extent of the Port Hills seismicity.

Stepwise eastwards progression of the first three thrust blocks exposes the deeper crustal structure in the intensively sheared Black Range-Harper block (Chamberlain et al. 1996), through increasingly greater preservation of cover sediments in the Craigieburn-Castle Hill block out to the Malvern Hills and the plains (Fig. 1; Forsyth et al. 2008). All of the three preceding areas expose pre-Quaternary geology that permits detailed geological mapping, complemented in the Malvern Hills case by seismic reflection lines on either side of the Waimakariri River at Springfield (Dorn et al. 2010). The younger and less evolved the structure, the more important is the role of fault propagation folding relative to fault displacement in accommodating the shortening and total throw on each fault. Further east, folding is the only surface indication of emergent blind thrusts from beneath the young sedimentary cover of the plains.

To the southwest, the thrust blocks appear to be truncated by a cryptic northwest-striking boundary structure followed by the Rakaia and Wilberforce Valleys (Group 7). To the northeast, similar thrust fault sheets spread from the Southern Alps with frontal thrusts at Mount Grey and the west side of the Culverden basin.

The pattern is complicated by the convergence of a west-facing coastal belt (Group 5), propagating onshore from the southern end of the Hikurangi subduction margin and extending westwards to the eastern side of the Culverden basin (Nicol et al. 1994). The opposing systems converge at Mount Grey (Fig. 1), but step back to the coast near the Ashley River. Along-strike southwards migration of this west-facing system onshore leads to coastal uplift, with growing hills closing the Waipara basin from the sea during Otiran glacial times and tectonically induced Holocene uplift and progradation extending south to at least the Ashley River mouth (Yousif 1985; Nicol et al. 1994; Al-Daghastani & Campbell 1995; Shulmeister & Kirk 1996; Dodson 2009). The progradation cycles identified by Shulmeister & Kirk (1996) are close to the dates for major range-front earthquakes inland (Howard et al. 2005) and to episodic down-cutting on the Waipara River, so may be a consequence of coseismic uplift. This temporal association suggests that pulses of eastwards migration of uplift and fault activation triggered by a major earthquake in the hinterland are characteristic of North Canterbury tectonics. The combined effect of the two opposing thrust systems has generated rising anticlines to form the Cust downlands and the hills surrounding the Waipara basin, exposing analogues of the structures still buried beneath the outwash gravels of the Canterbury Plains.

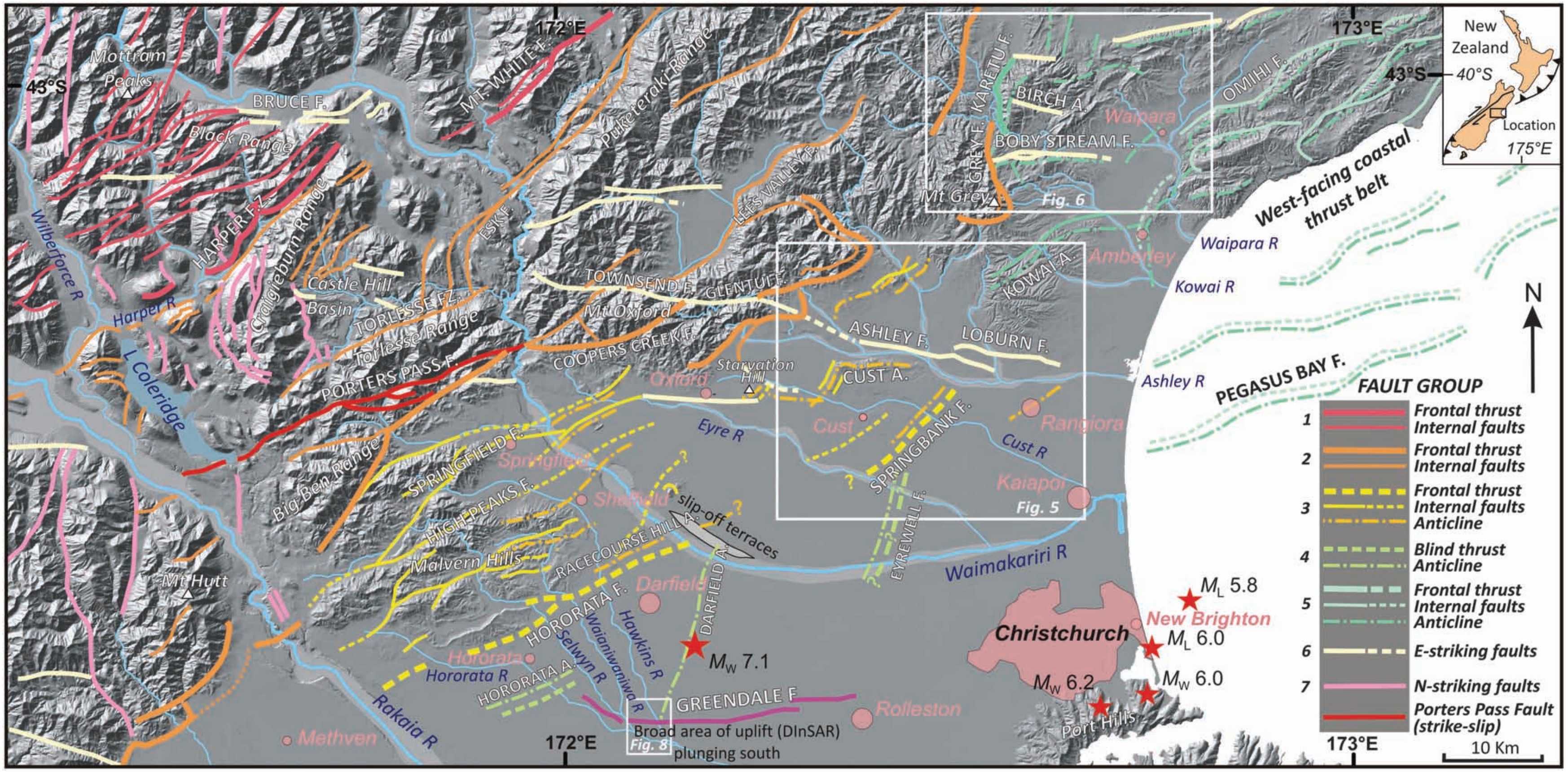


Figure 1 Map of the north Canterbury region showing main fault groups (colour coded). Dashed lines represent faults concealed underneath Quaternary gravels or offshore. Geographic localities referred to in text and the location of Figs 5, 6 and 8 are also shown. The location of the epicentres triggering cycles of seismicity are shown by red stars. F, Fault; FZ, Fault zone; A, Anticline; R, River. Inset shows location of Fig. 1 with respect to the current plate boundary across New Zealand.

Evolution of blind thrusts

The repetitive dimensions of the thrust block groups aligned with the onshore projection of the Chatham Rise basement high, the anomalously short rupture length to displacement ratios of the Greendale and Port Hills faults and the aftershock clusters oblique to these two faults suggest there are underlying controls in common. We propose that the strike-slip Greendale Fault is an element in a more extensive thrust-dominated pre-emergent structure. It is therefore worth considering how growth of a new thrust system will affect surface deformation and the distribution of internal strain within the evolving structure.

To illustrate these points, Trishear™ proprietary software (Allmendinger 1998) has been used to generate a simple generic model with a 10-km-thick basement overlain by 2 km of sedimentary cover, scaled to emulate the local depth to the brittle-ductile transition and the basal decollement. Other parameters are selected to reproduce a model that is geometrically similar to local examples. In the Canterbury setting, it is helpful to visualise shortening as a block of upper crust being carried west by basal viscous shear at the interface with the westwards advance of the lower crust. This block impinges against, and underthrusts, the backstop formed by the next-westernmost block to generate the relative overthrust geometry of the model.

Steps 1–5 in Fig. 2 show the incremental propagation of a listric thrust cutting up from the ductile shear zone of the basal decollement. The thrust ramp angle is assumed to be initiated at a low dip in the visco-elastic transition zone, but observed to break the surface at an angle as steep as 50–60°. Attention is drawn to the following points.

1. The minimum spacing between thrusts is constrained by the space needed for the overthrust slab to bend and pass up to the surface on the ramp formed by the frontal thrust. This distance is geometrically the apparent horizontal thickness (width) of the slab determined by the true slab thickness and the ramp angle, so that there is a simple relationship between increase in slab thickness and increase in thrust spacing. Material properties can modify this relationship by affecting the ramp angle and internal strain

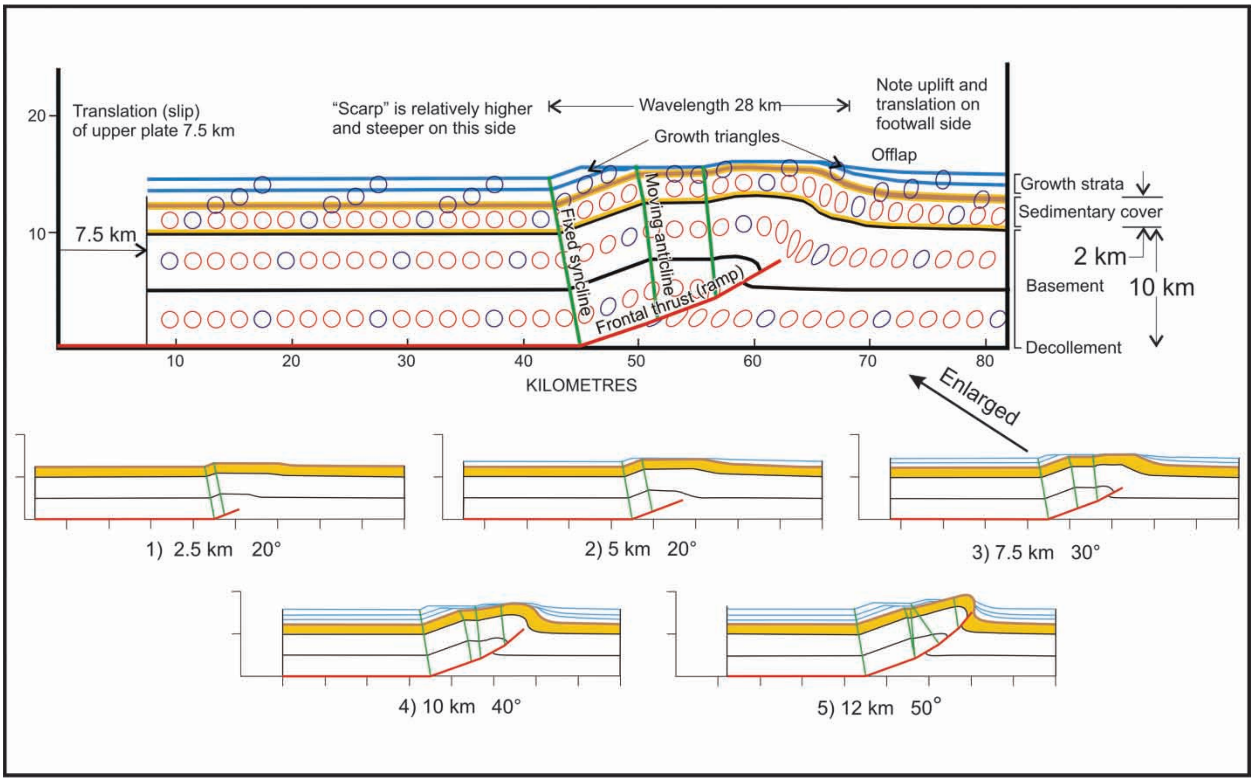
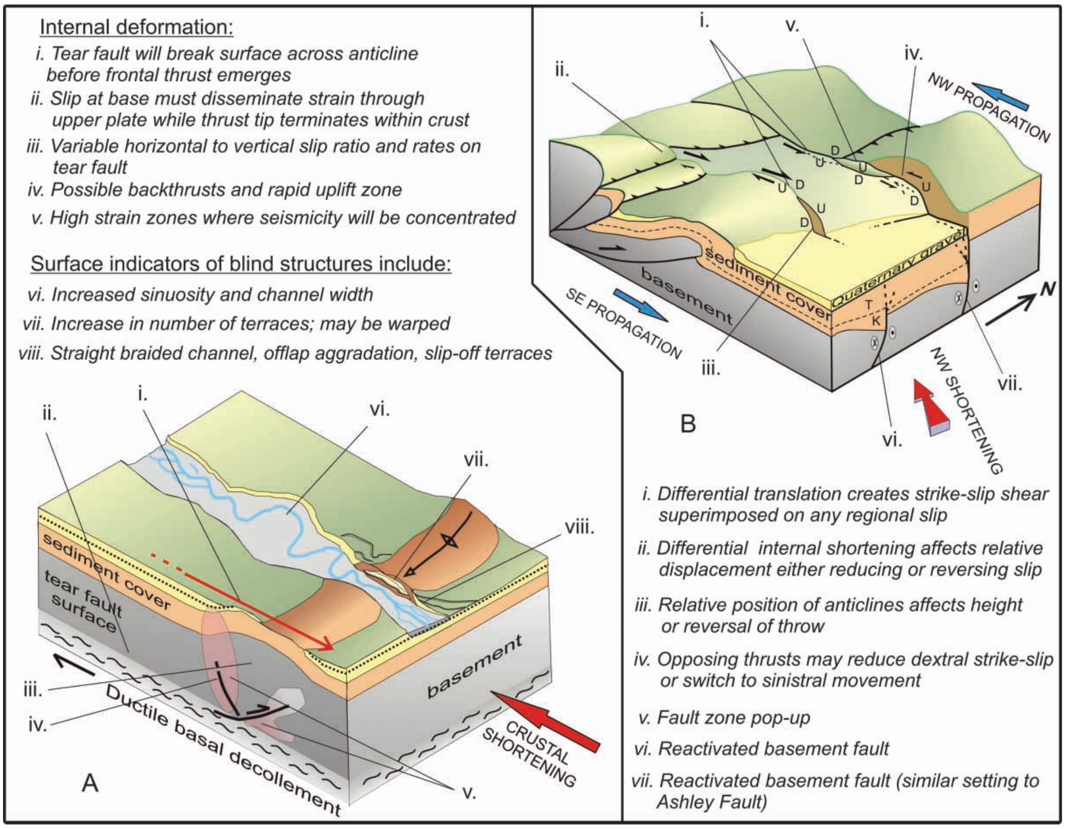


Figure 2 Trishear™ proprietary software was used to model a blind thrust and associated fault propagation, scaled to the upper plate crustal dimensions of Canterbury. The sequence from initiation to near emergence of the fault tip at the surface is depicted in steps 1–5. With each step, the increment of translation slip of the upper plate is shown in kilometres and steepening of the frontal thrust dip is shown in degrees. An intermediate stage (step 3) is enlarged and derived using the following parameters: slip = 7.5 km, propagation = 19 km, tip ramp angle = 30°, trishear angle = 60°, propagation to slip (P/S) ratio = 2.5. Strain ellipse markers show zones of high strain which are likely locations of secondary faulting or ductile deformation. Note the changes in distribution of surface deformation that will affect fluvial processes. In all steps, yellow and brown lines represent pre-deformation sedimentary cover (above basement) and blue lines represent growth strata during deformation. Scale ticks represent 10 km intervals.

(Bird 2003), but for a uniform depth to decollement and consistent basement lithology, a regularity in the repetition of thrust spacing is to be expected.

2. Folding involves non-recoverable strain accumulating over long periods of time. Aseismic creep is continually taking place, creating viscous shear at the basal decollement and convergence compression in the upper crust. While the thrust tip is still blind, this accumulating strain is largely dispersed through rock above the thrust. The slab must deform to pass through the fixed hinge zone at the point where the thrust breaks upwards from the base, and also to create fault propagation folds over the thrust tip (Fig. 2). In cold indurated rock such as typical Torlesse sandstone, this internal deformation likely involves quasi-plastic cataclasis on all scales of dislocation, as seen in deep basement exposures (Chamberlain et al. 1996). Precursor fold growth (Nicol & Campbell 2001; Campbell et al. 2003), ongoing background micro-seismicity and a long period of dispersed aftershocks following major coseismic slip on the main faults are likely consequences of this prolonged internal strain adjustment.
3. Initially, uplift and backtilting some 20 km behind the blind thrust front may be much more evident in its impact on ground surface deformation and on antecedent river bedforms than the more diffuse frontal thrust-related surface deformation (steps 1 and 2, Fig. 2). As the fold grows, incision and bedform changes will affect the course of a river over the whole rising anticline (Fig. 3A).
4. Backthrusts nucleating in the zones of complex strain near the base of the thrust, or created at changes in dip of the fault plane, can modify a simple translation of the slab up the ramp.
5. At the leading edge, uplift extends into the future footwall and the frontal scarp amplitude remains lower relative to the elevation change at the trailing edge, until the underlying footwall syncline has become established. Topographic indicators of the leading edge therefore may be less evident at the early stages (see enlarged detail of step 3, Fig. 2). As the blind thrust approaches the surface, a more obvious increase in amplitude and less box-like leading-edge anticline and footwall syncline pair evolves (steps 4 and 5, Fig. 2).



Internal deformation:

i. Tear fault will break surface across anticline before frontal thrust emerges
ii. Slip at base must disseminate strain through upper plate while thrust tip terminates within crust
iii. Variable horizontal to vertical slip ratio and rates on tear fault
iv. Possible backthrusts and rapid uplift zone
v. High strain zones where seismicity will be concentrated

Surface indicators of blind structures include:

vi. Increased sinuosity and channel width
vii. Increase in number of terraces; may be warped
viii. Straight braided channel, offlap aggradation, slip-off terraces

i. Differential translation creates strike-slip shear superimposed on any regional slip
ii. Differential internal shortening affects relative displacement either reducing or reversing slip
iii. Relative position of anticlines affects height or reversal of throw
iv. Opposing thrusts may reduce dextral strike-slip or switch to sinistral movement
v. Fault zone pop-up
vi. Reactivated basement fault
vii. Reactivated basement fault (similar setting to Ashley Fault)

Figure 3 A, Block diagram showing the internal distribution of strain associated with a blind thrust and the surface response of antecedent drainage and landforms. The left-hand face of the block diagram represents a tear fault surface ideally perpendicular to the blind thrust. **B,** Block diagram showing east-striking reactivated basement faults intersecting a northeast-striking thrust fault and fold system. This shows schematically how relative rates of translation and the distribution of internal shortening by folding can modify the slip distribution along the east-striking reactivated faults. The size of the movement arrows on the east-striking faults is proportional to the amount of relative slip. T, Tertiary rocks; K, Cretaceous rocks; U, Up; D, Down.

6. The relative proportion of the total throw distributed between fold amplitude and fault displacement is a measure of the importance of internal ductile deformation and plasticity relative to brittle failure on the main fault. By setting the propagation to slip (P/S) ratio to 2.5, being the rate of propagation of the thrust fault tip relative to basal-slip displacement onto the ramp, folds comparable in profile to local examples of anticlines and footwall synclines are produced.
7. Although non-recoverable strain accounts for much of the internal deformation shown by the strain ellipses in Fig. 2, there will be a component of stored elastic strain. This is because an effective plastic strength is imposed by intragranular strain hardening and frictional resistance to shearing in cataclastic deformation. Release of this elastic strain in the wall rocks would contribute to the energy drop on fault rupture, either as part of the thrust system or of any independent transecting fault.

Strike-slip faults

Strike-slip faults in Canterbury are also present and, because many show obvious active traces, they have been the subject of several palaeoseismic investigations. The Porters Pass–Amberley Fault Zone (PPAFZ) is probably the best known (Cowan 1992; Howard et al. 2005) and is inferred to be at the early stages of evolving into the newest member of the southward-migrating strike-slip Marlborough Fault System (Cowan et al. 1996). In the west, the Porters Pass Fault segment cuts obliquely across the Torlesse Range (Fig. 1) and records discrete transpressive strike-slip. East of the Waimakariri River, the PPAFZ as a whole accommodates dextral transpressive shear by stepping through and exploiting pre-existing faults along the Mount Oxford to Mount Grey range front (Fig. 1). From Mount Grey to the Waipara River mouth, the PPAFZ passes into the northeast-trending coastal folds which are locally rotated clockwise into an en echelon, sigmoidal pattern (Yousif 1985; Nicol et al. 1994; Al-Daghastani & Campbell 1995). Cowan et al. (1996) also incorporated the Lees Valley and Ashley faults as PPAFZ splays.

Here we focus on broadly east-striking faults scattered throughout Canterbury because of their relevance to the similarly oriented Greendale and Port Hills faults (Group 6 faults in Fig. 1). Examples include the Glentui and Townsend faults north of Mount Oxford, Boby Stream Fault, Ashley Fault and those as far inland as the east-trending reach of the upper Waimakariri River such as the post-glacial Bruce Fault trace (Chamberlain et al. 1996). Where sense is demonstrable by offset topography (Nicol & Campbell 2001) and right-stepping shear folds (Yousif 1985; Litchfield et al. 2003), these faults have been found to be dextral transpressive oblique-slip structures. Importantly, these faults only have a mappable surface expression of young displacement where they are in close association with faults and folds of the northeast thrust system. They truncate folds and segment thrusts, linking steps in the thrust front, but terminate as visible traces close to the projected limits of the folds and thrusts.

The association of these east-striking faults with reactivated and inverted Cretaceous normal faults in the basement is evident in seismic reflection lines (Ghisetti & Sibson 2012; Jongens et al. 2012). Ghisetti & Sibson (2012) discuss the reactivation in terms of the regional principal compressive stress orientation quoted as $115 \pm 5°$, which is broadly perpendicular to the northeast-striking thrust system. The east-striking anisotropic planes of weakness are optimally oriented for strike-slip movement (Sibson et al. 2011) but have non-vertical, possibly listric, dips in the basement (Ghisetti & Sibson 2012). If this regional stress were the only controlling factor, then these faults should propagate to the surface independently of other structures. Instead, these fault planes are preferentially reactivated in close association with the more dominant northeast-striking thrust faults and propagation folds, indicating that additional controls are involved. As will be shown below, these faults do not record pure strike-slip movement but rather dextral transpressive slip with variable throw.

A different sort of strike-slip faulting can take place in association with thrust faulting. Tear strike-slip faults commonly develop subperpendicular to the thrust front i.e. parallel to σ_1, the maximum principal compressive stress. They are commonly short but can extend well back into the upper thrust sheet, allowing thrust segments to advance independently. Rather than being solely controlled by the regional stress, the strike-slip shear on tear faults accommodates the differences in the amount of shortening being taken up in each thrust segment. Consequently, the tear faults must break to the surface early to allow for segmentation, differential translation and folding, but do not extend beyond the thrust front. If optimally oriented transecting faults exist already, then these are likely to be preferentially exploited as tears to take up the differential movement; they may be relatively quiescent beyond the active thrust environment (Fig. 3B), however. Juxtaposition of different parts of neighbouring thrust segments will create variations in net relative displacement and slip orientation along the length of the transecting faults (Fig. 3B). They may also be overprinted by the distribution of thrust-related strain in the wall rocks, producing anomalies in the uniformity of coseismic slip data (Fig. 3A).

In North Canterbury, east-striking transpressive faults are seen to generate associated anticlines parallel to their strike along the upthrown side. Consequently, where these intersect the northeast-trending folds, characteristic cross-folded interference structures result in distinctive L-shaped anticlines and triangular domes (Fig. 4). These include the L-shaped Cust Anticline (Fig. 5), and range from the fundamental basement block scale of the Doctors Dome (Fig. 6; Nicol 1993b) to second-order faulting and folding within the larger blocks such as the Onepunga Anticline

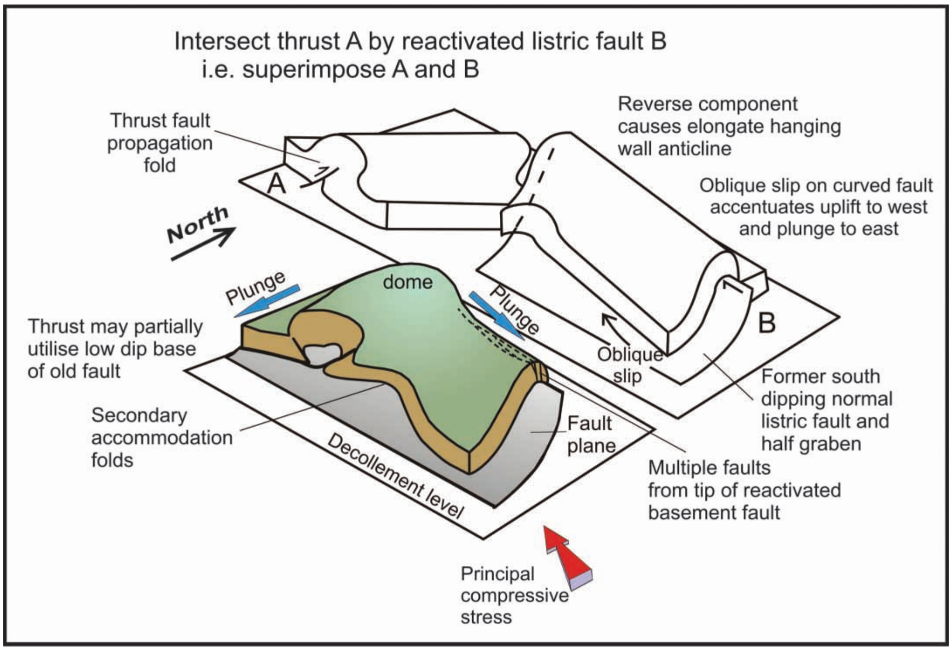


Figure 4 Block diagram demonstrating the effect of a transpressively reactivated south-dipping listric fault with an associated hanging-wall anticline, intersecting a northeast-trending thrust-propagated fold structure. The resulting superimposed folding includes L-shaped anticlines and triangular domes.

(Fig. 6) and at Starvation Hill (Fig. 1; May 2004). North of the Waimakariri River, elements of both the thrust and strike-slip systems combine into a succession of similar interacting structures, sequentially more developed and deeply exposed towards the foothills. These structures are thought to be possible analogues of the structural style now evolving south of the river under the gravels of the Canterbury Plains, and are described from south to north in the following three sections.

Cust Anticline, Ashley and Springbank faults

The Cust Anticline is topographically the most prominent feature of the northern plains. It is a long ridge sloping up to the west to culminate 150 m above the south side of the Ashley River (west of Mairaki Downs) before turning to the south, where it is breached by a broad saddle that is eroded and crossed by the antecedent Cust River (Fig. 5). North of the Ashley River, the surface scarp of the Ashley Fault can be traced parallel to the length of the Cust Anticline ridge. Upthrown on the south side, it reverses throw east of the Okuku River where uplift is on the north side. The ridge itself is underlain by a flower structure of faults over older Cretaceous structures (Ghisetti & Sibson 2012; Jongens et al. 2012).

South of this ridge and between the Cust and Eyre rivers, there is a lower loess-covered surface known as the Cust downlands which is elevated to about 50 m above the post Last Glacial Maximum (LGM) degraded aggradation gravel (Springston Formation) surface. The morphology of this higher surface has been shown from seismic reflection lines to match a northeast-trending anticline on the hanging wall of the southeast-facing blind Springbank Fault (Jongens et al. 1999, 2012). In a higher-resolution line undertaken by Estrada (2003) alongside the Cust River, the fault tip is observed to terminate within the Pliocene–Pleistocene Kowai Formation but close to the point of reaching the surface beneath a growth triangle of Pleistocene aggradation gravels.

The Springbank Fault appears to die out to the south, but a short seismic reflection line immediately north of the Eyre River revealed the crest of another gentle anticline (Figs 1, 5). This structure was interpreted by Estrada to indicate an associated blind fault, named the Eyrewell Fault, and located near a steep gradient anomaly on the Waimakariri River topographic profile (see Fig. 7). Although unidentified topographically south of the river, some Darfield earthquake aftershocks extend towards the Eyrewell Fault from the east end of the Greendale Fault.

Post-100 kyr fold growth of the Cust Anticline has exposed the Pliocene–Pleistocene Kowai Formation where once a combined Cust and Ashley River occupied a channel over this surface (Estrada 2003). Drainage was largely diverted to the present course of the Ashley River by a backthrust forming the NNE-trending western end of the Cust Anticline (Fig. 5). The east–west extent of the elevated structure is only about 12 km, but the Springbank Fault lies approximately 20 km east of the main range-front fault north of Oxford. The model in Fig. 2 step 5 indicates that growth of the fault propagation fold will move the zone of maximum uplift towards the frontal thrust as the fault tip approaches the surface.

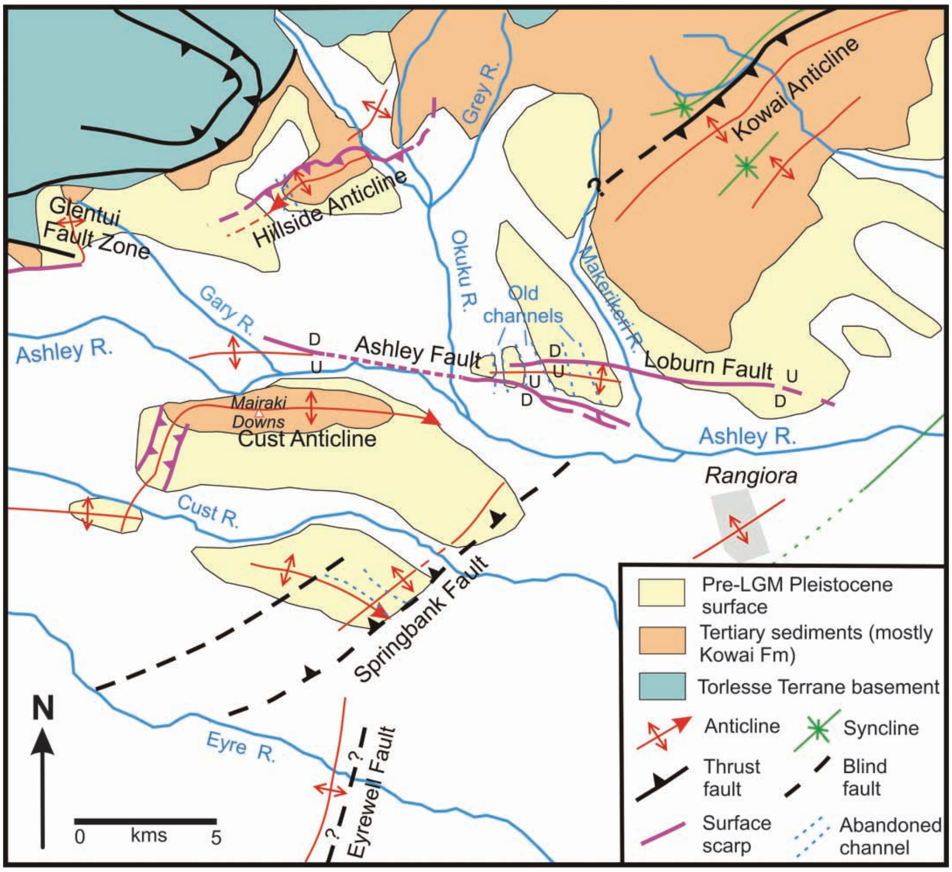


Figure 5 Simplified geological map of the Ashley River area west of Rangiora, showing the Cust Anticline, Springbank Fault, Ashley Fault and other structures described in the text. Location of map is shown on Fig. 1. Map adapted from Sisson et al. (2001), Estrada (2003) and Forysth et al. (2008).

The associated strike-slip faults, both before and after the thrust tip breaks the ground surface, can document the activity on the thrust system. This is well illustrated by the Ashley Fault trace where post-glacial rejuvenation of the scarp evidently propagated eastward, decreasing in scarp height across a mid-Holocene abandoned channel of the Okuku River to terminate at a point coinciding with the projected strike of the Springbank Fault (Sisson et al. 2001). This incremental rejuvenation of the Ashley Fault tip may be tracking emergence of the thrust fault. Discrete increments of propagation are likely to coincide with seismic events and these may also involve the growth of the adjacent folds. A date of 4280 ± 179 yr BP and possibly a second date of between 3500 and 3200 yr BP are indicative of the last identified displacement(s) on the Ashley Fault (Sisson et al. 2001). These dates are compatible with the broad estimate of c. 3000 yr recurrence intervals for the Springbank Fault where a low slip rate of 0.22 mm/yr has been derived from the deformation geometry and age of abandoned late Pleistocene surfaces (Estrada 2003). Estrada estimated a maximum moment magnitude of 6.4, assuming that slip is constrained to the Springbank fault plane only.

Immediately to the north of the Ashley Fault there is a complication, as the Ashley River appears to mark the inland southern boundary of the west-facing eastern thrust system (Group 5) before it steps back to the coast. Kowai Formation gravels are further elevated on west-facing thrusts and folds in the hills east of the Makerikeri River (Fig. 5). The boundary is defined by the seawards continuation of the east-striking inherited Cretaceous basement faulting at depth influencing not only the Ashley Fault, but also the similar Loburn fault. The latter fault has propagated from the east as determined by the chronology of intersected abandoned river channels (Sisson et al. 2001), probably responding to the west-facing thrust translation, to overlap the tip of the Ashley Fault. An anticline forms a pop-up ridge at the overlap, locally reversing the sense of throw on both faults (Fig. 5). Feature 'vii' in Fig. 3B schematically represents the structural setting of the Ashley Fault.

Onepunga Anticline and Boby Stream Fault

Structures in the Middle Waipara River valley may be the best analogue for the future evolution of the Greendale–Hororata fault system. Here faults have large surface displacements and are eroded to the level of the interface between basement and cover rock (Fig. 6). Torlesse Group basement is exposed in the culminating dome of the Onepunga Anticline. It too has a long anticlinal ridge plunging to the east, where the north limb is cut by the south-dipping dextrally transpressive Boby Stream Fault. Several strands splay into the cover sequence to enclose a faulted duplex, progressively shearing out the steep north limb. Boby Stream and the Waipara River cross the fault in several places and provide deeply incised sections through the duplex structure, exposing details of the disruption to the Cretaceous and Tertiary cover stratigraphy. Late Quaternary displacement has been variously partitioned across these strands as 1–2-m-high scarps across intervening terraces.

The Boby Stream Fault illustrates the ambiguities that arise in using the strike separation of marker units to determine the amount and sense of slip on dipping transpressive faults. At 'The Deans' (Fig. 6), there is a dextral offset of terraces displaced by the last event c. 300 years ago (Nicol & Campbell 2001). However, on the larger scale of mapped Tertiary units, offsets appear to contradict the shear sense of the terrace data. The Oligocene Limestone dips uniformly southeast at 30° on the north side of the Waipara River but is apparently separated sinistrally by 5 km across the Boby Stream Fault (Fig. 6). This is an artefact of the presence of a well-developed footwall syncline, now obliquely overridden.

The culminating dome of the Onepunga Anticline is created by a cross-cutting fold, but no associated fault crops out. Holocene deformation on this fold is evident from upwards warping of a portion of the 2300-year-old terrace on the flank. Westward, the end of the Boby Stream Fault disappears below the lobe of the east-facing Grey thrust fault which has overridden the cover sediment sequence (Fig. 6).

The Boby Stream Fault has ruptured three times in the last 2300 years, most recently approximately 300 years ago when the Grey and Karetu thrust faults, plus the Birch Fault, were all activated simultaneously as far as could be determined within the limits of dating (Nicol & Campbell 2001). This emphasises the intimate mechanical connection between thrust and strike-slip systems. Furthermore, the event was preceded by about 200 years of accelerated 20–25 m localised river downcutting in addition to regional incision, and attributed to precursor fold growth (Nicol & Campbell 2001; Campbell et al. 2003). Displacements of c. 5 m dextral and 1 m reverse recorded on a 300-year-old scarp of the Boby Stream Fault at 'The Deans' are similar to those on the Greendale Fault (Quigley et al. 2012). The apparent surface rupture length of 9 km is even shorter than the Greendale Fault, although the fault extends an unknown distance beneath the Grey thrust fault and eastwards as a blind structure into the Waipara Basin (Dodson 2009). This last event and other dated events on the Porters Pass Fault at 500–600 and 800–1100 yr BP (Howard et al. 2005) imply that the 2010–2012 Canterbury earthquake sequence is the

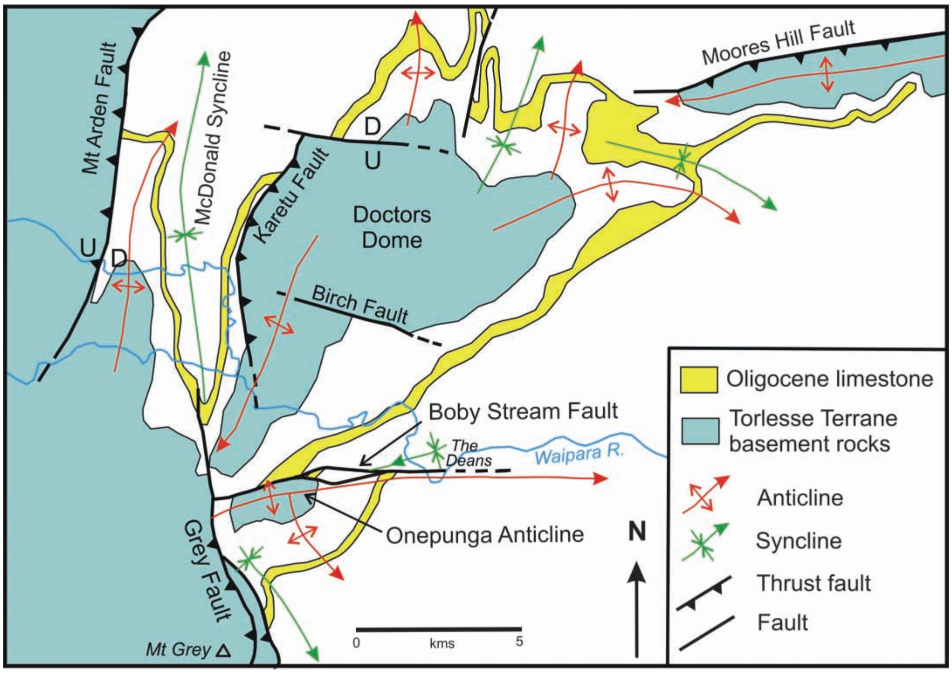


Figure 6 Simplified geological map of the middle reaches of the Waipara River showing the Grey Fault, Boby Stream Fault, Doctors Dome, Karetu Fault, Onepunga Anticline and other structures described in the text. Location of map is shown on Fig. 1. Map modified from Nicol (1993b).

fourth recognised event likely to have exceeded M_W 7.0 within a potentially damaging 50 km radius of Christchurch at c. 300 year intervals.

The Waipara palaeoseismic evidence for near-synchronous interaction at the junction of the major thrusts on Grey and Karetu faults with strike-slip movement on the east-striking Boby Stream and Birch faults reinforces the view that coseismic rupture of the east-striking faults is intimately associated with thrust faulting. The strike-slip displacement, at least in part, may be accommodating differential thrust fault displacement and shortening within the adjacent wall rocks. The resulting rupture surface will be a complex interconnected network of non-planar faults, substantially greater in area than that suggested by the length of the surface fault rupture.

Doctors Dome and Karetu Fault

North of the Boby Stream Fault, the triangular area of exposed Torlesse basement known as Doctors Dome represents the deepest exposure level of the serial sections through this structural style (Fig. 6). Mapped by Nicol (1993b) and in part by Litchfield et al. (2003), the structure is bound on the west by the Karetu Fault and partly overrides the east limb of the tight McDonald Syncline that is sandwiched against opposing thrusts (Fig. 6). At the south end of the Dome, the Karetu Fault appears to terminate and strands of the Grey Fault cut obliquely across the McDonald Syncline. Around the north and east sides of the Dome, larger folds in the surrounding cover rocks (traced by the Oligocene limestone) are cored by deformed Torlesse basement with no evidence of detachment at the unconformity. East-striking faults cut across the Dome and there is a complex interplay of north- and west-striking faults at the north-western end. The prevalence of folding on these two trends around the margins of the structure, including cross-folding down the length of the McDonald Syncline, can only be hinted at by the outcrop pattern of the marker Oligocene limestone illustrated at the scale of Fig. 6, but is discussed in detail by Nicol (1993b). Many of these smaller-scale folds are not obviously fault-related, appearing to be secondary fold adjustments to the complex geometry and strains involved in basin and dome cross-folding. The net result is shortening in two directions being taken up within the basement as well as the cover rocks.

Subtle ground deformation on the Canterbury Plains

Hidden structures, potentially analogous to those exposed in North Canterbury, are thought to be evolving under the Canterbury Plains gravels and are represented at the surface by subtle geomorphic indicators of active ground deformation. Small rivers are particularly sensitive to ground deformation, responding by changes in gradient and bedform. Even the large gravel-bed rivers, which would be expected to maintain grade, show signs of bedform response to ground deformation. By quantifying channel characteristics and accurately surveying reference surfaces such as terraces, substantial progress has been made in identifying subtle deformation features and providing constraints on the timing and rates of deformation on more evolved structures (e.g. Yousif 1985; Estrada 2003; Litchfield et al. 2003; May 2004; Duffy 2008). Often simple air photo interpretation will identify potential target sites, such as the response of antecedent rivers to growing folds (summarised in Fig. 3A).

A typical small-scale example occurs where the Hawkins River crosses the eastern end of the Greendale Fault to join the Selwyn River (Figs 1, 8). Incision into the late LGM gravels indicates a pre-Darfield earthquake history of arching in the Holocene. Along the Hawkins River, there is a typical association of increasing sinuosity and lateral planation as the river gradient decreases due to incremental backtilting on the upstream side of the anticlinal uplift crest. This is followed by the onset of incision and trapping of the river channel, and progressive narrowing of the course at the crest. On crossing the rising structure, the steepening gradient causes a change to a straight braided course down to the confluence with the Selwyn River. Emergence of the new Greendale fault scarp occurred at a sharp bend in the channel on the upstream side of the anticline. Slip-off terraces where the Selwyn River flows parallel to the south side of the rise may be indicative of periodicity in the uplift of a propagation fold over the emerging fault tip (Fig. 8). The timing of the onset of folding is not known until the terraces can be dated, but it was clearly coeval with the time taken to incise the Selwyn River.

Range-front structures near Hororata show plenty of evidence of post-glacial and Holocene activity. Particularly significant is the Hororata anticline (Fig. 1), an actively growing fold trending southwest towards the Rakaia River and expressed as raised downland. The anticline is associated with an imbricate thrust splaying from the footwall of the frontal Hororata Fault (Jongens et al. 2012). The course of the Hororata River upstream of the anticline is parallel to the range front before turning sharply southeast, to align with the Greendale Fault western segment in the lower reaches. A former course of the river was incised into the anticline and has now avulsed off the rising flank into the present course.

Even more recently, the river has formed a second, now abandoned, channel at the point where it bends southeast. The anticline was uplifted on the 4 September 2010 Darfield earthquake (Beavan et al. 2010), accompanied regionally by strong vertical acceleration. A modelled M_W 6.5 rupture occurred near the intersection with the western segment of the Greendale Fault (Gledhill et al. 2011). Aftershock activity is concentrated along or near the anticline and consistently demonstrates reverse moment tensor solutions (Fig. 6 of Gledhill et al. 2011). This associated thrust, stepping forwards of the Hororata Fault, occupies a position relative to the Greendale Fault which is comparable to the position occupied by the Grey thrust lobe relative to the Boby Stream Fault.

Northeast along-strike of the Hororata Fault, similar channel changes occur on the Waianiwaniwa River on crossing the fault. Points of rapid bedform change, local incision and slip-off features are yet to be systematically investigated in rivers draining the range front. Less speculative is the location of the Racecourse Hill Anticline on the hanging wall of the fault beside State Highway 73 (Jongens et al. 2012). This is confirmed in seismic reflection lines (Finnemore 2004) to be an anticline uplifting and preserving older gravels from erosion during deposition of the LGM outwash gravel around the base of Racecourse Hill. The younger surface on these gravels is also visibly arched, as seen on the State Highway and adjacent railway. Re-levelling of benchmarks has shown detectable growth over a 40 year period after 1949 (A. Bal, pers. comm. 1996).

The DInSAR data of Beavan et al. (2010) show a discrete zone of anticlinal uplift which coincides with ongoing aftershock activity just west of the Charing Cross epicentre. This is interpreted by Beavan et al. as a blind thrust dipping southeast to the 10.8-km-deep hypocentre location. The zone of coseismic uplift coincides with a low topographic rise on Clintons Road; unless this is an erosional artefact, it is probably the product of several increments of post-glacial growth deforming the LGM aggradation surface. South-westwards, this anticline projects into the footwall of the Greendale Fault at the Hawkins River site (Fig. 8) and it may be no coincidence that this is the general location of the stepover to the western segment of the fault and the late Holocene deformation described above. The anticline at its northern end is picked out by a small cluster of aftershocks which swings northwards to pass just east of Darfield where there is a similar rise in the main highway. The zone extends to the Waimakariri River at the point where slip-off terraces occur on the north bank. The structure is referred to here as the Darfield Anticline.

The best evidence for the existence of anticlinal uplift on the larger scale of a fundamental thrust block, thought to be propagating eastwards from the Hororata thrust front, comes from a Waimakariri River stream gradient analysis by Estrada (Fig. 7; 2003). Marked gradient anomalies occur over the inner plains where major faults such as the Springfield and High Peak faults bring resistant basement to the surface, but the Hororata Fault is also picked out by a steepening grade (Fig. 7). Downstream, a zone of lower gradient coincides with the proposed back-tilted side of the Darfield Anticline and the slip-off terraces shown in Fig. 1. The second of two points with particularly steep Gradient Indices (GI) values of 4.3 and 4.7 lies 25 km downstream of the Hororata Fault (Fig. 7); the gradient changes over the entire 25-km-reach invite comparison with the surface profiles of steps 1 and 2 in Fig. 2. The downstream nick point marks a consistent change from degradation to aggradation as regularly monitored by decades of re-measurement records from the Canterbury Regional Council (Estrada 2003), aligned with the general location of Estrada's proposed blind Eyrewell Fault described above.

South of the Greendale Fault, evidence for southwards tilting is seen in the slip-off terraces on the Rakaia River north bank and as coseismic uplift (Beavan et al. 2010) showing a broad south-plunging arch (Fig. 1). The latter indicates that the same incipient broad anticlinal structure continues on the south side of the Greendale Fault.

Port Hills seismicity

Interpreting the Port Hills seismicity, including that immediately offshore from Christchurch, is complicated by the lack of surface rupture, the complexities of subsurface structures beneath the Banks Peninsula volcanoes and the

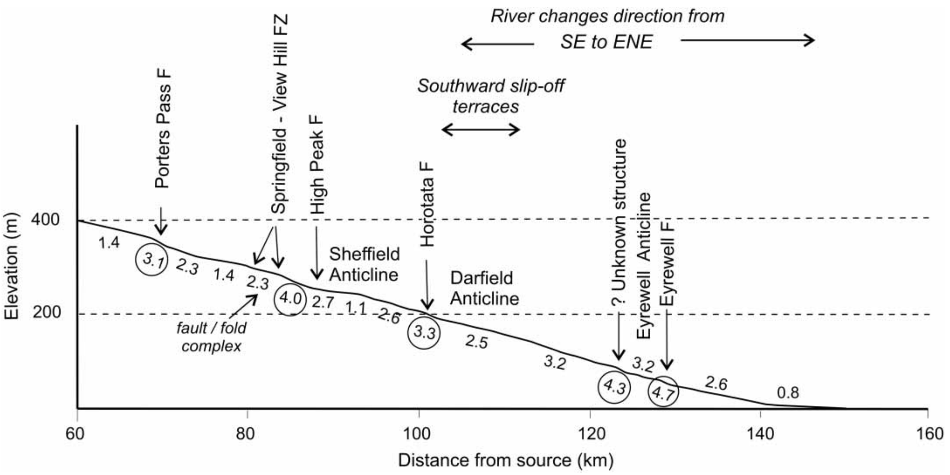


Figure 7 Waimakariri River profile with gradient indices. Circles mark nick points with anomalous steep gradient indices. Modified from Estrada (2003). F, Fault; FZ, Fault zone.

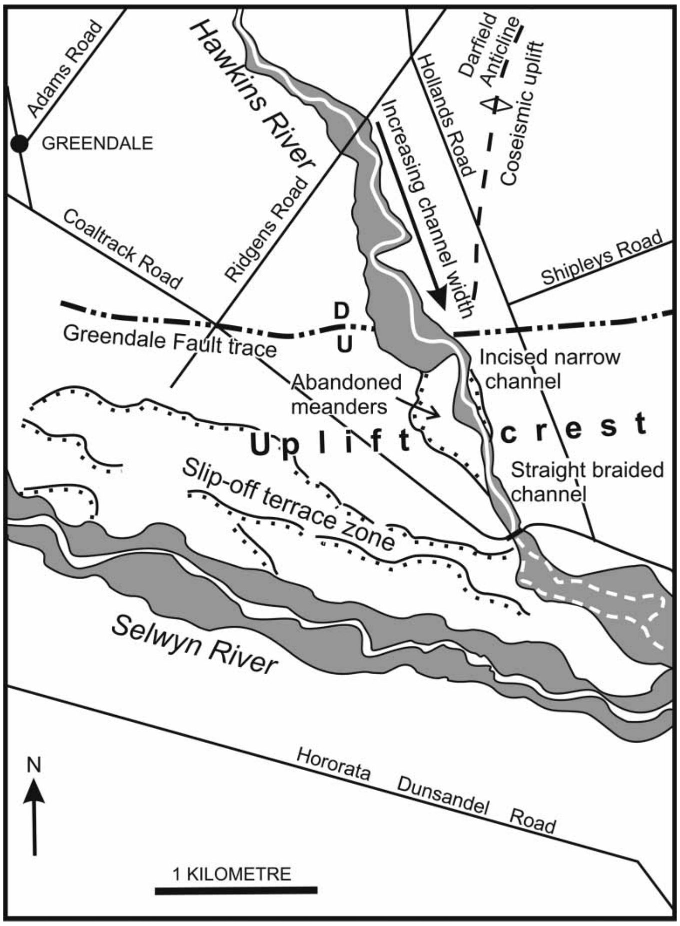


Figure 8 Sketch map of the Hawkins River where it crosses the Greendale Fault to join the Selwyn River. Geomorphic evidence of pre-2010 rupture Holocene uplift south of the Greendale Fault is indicated. Location of map is shown on Fig. 1. DInSAR measured uplift occurred on the southwest extension of the Darfield Anticline where shown, and on the anticlinal uplift crest south of the Greendale Fault at the position indicated by pre-earthquake river bedform changes.

stress perturbation introduced by topographic gravitational loading by the volcanoes. In relation to this paper, however, some points arise from the new geophysical data. The Port Hills seismic activity seems to be discretely separated and offset from the end of the Greendale Fault. The New Brighton 23 December 2011 aftershock sequence provides strong evidence for an offshore thrust striking southwest coming onshore to intersect the Port Hills fault (GeoNet 2011), of which the latter has accommodated a significant component of dextral strike-slip movement (Beavan et al. 2011; Kaiser et al. 2012). The thrust possibly extends under the south-dipping Port Hills fault footwall to intersect with the section activated by the 22 February 2011 M_W 6.2 earthquake, triggered by an oblique thrust focal mechanism. The modelled slip distribution (Beavan et al. 2011; Kaiser et al 2012) and the concentration of the larger-magnitude aftershocks at the eastern end of the Port Hills fault indicate it may be a westerly-propagating strike-slip rupture zone. The possibility that the Port Hills system is related to the west-facing coastal faults (Group 5) should be considered, and a parallel can perhaps be drawn between this system and the opposing faults in the Ashley–Loburn structural relationship described above.

Conclusions

The sequential propagation of both the thrust systems from the west and the progressively younger hybrid thrust and strike-slip assemblages from the north into the current seismically active area strongly indicates that similar processes are involved. Fundamentally, the deformation seems to be dominated by crustal shortening with an incipient regional component of dextral shear associated with the southern limit of the Marlborough Fault System. It is proposed here that the Greendale Fault may transect the early stages of a thrust system propagating southeast from the Hororata range-front faults in the early stages of a new cycle of upper crustal detachment.

The combination of tectonic geomorphic indicators of late Quaternary active deformation on the Canterbury Plains west of Christchurch and the distribution and characteristics of the 2010–2012 Canterbury earthquake sequence are compatible in scale and pattern with the adjacent, but better exposed, structures of the Canterbury foothills. Emergence of the Greendale Fault, indications of anticlinal growth from landscape features and geophysical evidence indicate that the 2010–2012 earthquake sequence relates to concurrent thrust and strike-slip fault activity on an evolving thrust system at depth. The triggering of three of the four cycles in the current earthquake sequence by $\geq M_W$ 6 events with northeast-striking nodal planes and reverse moment tensor solutions reinforces the primary role of thrust faulting, particularly as two triggering events (Charing Cross and New Brighton) are several kilometres north of the activated strike-slip faults.

The east-striking faults, derived from non-vertical inherited Cretaceous normal faults in the basement, are reactivated as dextral transpressive faults propagating through the Cenozoic cover sequence as steep faults evolving into multi-stranded flower and imbricated zones. The convergence component creates associated folds, expressed topographically as elongate anticlines rising up-plunge westwards as the hanging wall is thrust obliquely across the footwall. Where these faults intersect the northeast-striking folds and faults, characteristic cross-folded interference structures are created.

While these reactivated basement faults are subject to the regional stress field, the east-striking faults seem to be either entirely activated or selectively accelerated to the surface in association with the major fault propagation folds of the thrust system. They will tend to emerge as those folds appear, while the thrust faults may still be blind. Consequently, the faults will be prominent surface features in the early development of this combination of structures and can record closely interlinked palaeoseismic activity on both sets of structures.

The length of the surface rupture on the Greendale Fault and other analogous faults may not be a good measure of the fault plane surface area involved. This is because the surface area may be increased if the fault is listric, includes multiple strands or if rupture involves synchronous interaction with adjacent thrust faults. Further, variations in local net-slip magnitude and orientation along the rupture trace may relate to overprinting by thrust-related relative translation. The stored elastic strain associated with internal shortening of the thrust-related folding may also contribute to the observed distribution of slip and the total energy release during the main and aftershock sequence.

The Greendale Fault with its associated seismicity pattern and ground deformation seems compatible with the proposal that it is the secondary product of the next increment of a crustal thrust block detachment propagating east from the Hororata Fault front. The Port Hills system, on a similar scale, shows evidence in terms of seismicity and coseismic deformation of an association with active thrust faulting.

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