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Erosional control on the structural evolution of a transpressional thrust complex on the Alpine Fault, New Zealand

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Abstract—The Waikukupa thrust is a 4-km long oblique thrust segment of the New Zealand Alpine Fault that has developed over the last 65,000 years, emplacing mylonite and cataclasite over fluvioglacial gravel. During the last 20,000 years or so, a 2-km long 'out-of-sequence' imbricate, the Hare Mare thrust, has formed on the east side of the Waikukupa River valley. The average rate of slip on the two thrusts is estimated at 22–30 mm/year. Analysis of the fault structure in terms of critical wedge theory is consistent with imbrication resulting from a reduction in wedge taper below the critical value due to rapid river erosion. The internal structure of the Waikukupa thrust sheet consists of reverse faults and angular folds forming duplex-like systems, associated with minor strike-slip faults. Within the limits of uncertainty, the structures are consistent with strain accumulating during internal transpressional shear of the thrust sheet while the taper of the latter was subcritical prior to the transfer of fault displacement to the Hare Mare imbricate. The thrust complex forms part of a serrated range front fault system consisting of oblique thrust sections linked by strike-slip faults. We introduce the term 'serial partitioning' for this type of partitioned transpressional fault system. Its development is linked to the erosional processes at the range front. An alternative system consisting of parallel thrust and strike-slip faults we here term 'parallel partitioning'. While angle of obliquity is an important parameter in determining which partitioning model develops, we suggest that erosion rate is also an influential factor. © 1997 Elsevier Science Ltd.

INTRODUCTION

Plate boundaries have traditionally been divided into divergent, transform and convergent end-members. In recent years, increasing attention has been paid to the importance of oblique motion leading to components of transtension and transpression (Harland, 1971) along the boundary. In particular, orogenic belts both recent and ancient are increasingly recognized as commonly having a transpressional character (e.g. Ratschbacher, 1986; Brown and Talbot, 1989; Hansen, 1989; Oldow *et al.*, 1990; Cobbold *et al.*, 1991 Holdsworth and Strachan, 1991). The Southern Alps of New Zealand, bounded by the Alpine Fault, form one of the world's most accessible, active transpressional plate boundaries (Walcott, 1978a; Norris *et al.*, 1990). The current plate vector may be calculated from global plate motions (DeMets *et al.*, 1990) as being 15–20° oblique to the average strike of the Alpine Fault within South Island, although preliminary calculations based on recent GPS data suggest a higher angle than this for present-day motion (Walcott, 1996).

Understanding of the mechanics of deformation of thrust belts has increased considerably following the development of critical wedge models of convergence (Chapple, 1978; Davis *et al.*, 1983; Dahlen, 1984). Koons (1990) showed how continental collision would produce a two-sided critical wedge, and how efficient erosion on one side of the wedge would result in an asymmetry with a high proportion of displacement concentrated on the eroding margin. This was also shown in an elegant numerical model by Beaumont *et al.* (1992). Preliminary mathematical models of three-dimensional critical wedges, where movement is oblique to the indentor, have been published by Platt (1993) and Koons (1994).

Transpressional zones have been modelled in terms of homogeneous strain by Sanderson and Marchini (1984), Fossen and Tikoff (1993) and Tikoff and Teyssier (1994), although Robin and Cruden (1994) have developed a modified model in which strain is inhomogeneous. Such models are useful in understanding finite strain but are limited in understanding the mechanics by the condition of homogeneous strain and by their effective two-dimensional nature which precludes topography and erosion. In some zones of oblique convergence, a partitioning of displacement onto separate parallel structures accommodating strike-slip and thrusting has been described (Fitch, 1972; Walcott, 1978b; Mount and Suppe, 1987; Namson and Davis, 1988, 1990; McCaffrey, 1991). Displacement partitioning of this nature has been the subject of much discussion (Zoback *et al.*, 1987; Cobbold *et al.*, 1991; Molnar, 1992; Platt, 1993). The central section of the Alpine Fault, on the other hand, shows little discernable partitioning of this sort, accumulating both dextral strike-slip and convergence in a roughly similar ratio as that of the plate vector (Bibby *et al.*, 1986; Norris *et al.*, 1990; Berryman *et al.*, 1992). At shallow depths, however, the fault trace appears to develop an en échelon character of oblique thrusts linked by strike-slip faults on a scale of 1–5 km (Norris and Cooper, 1995).

Development of oblique thrusting on a section of the San Andreas Fault in southern California was described by Sylvester and Smith (1976) and interpreted as due to transpression across the fault. More evolved thrust development along a transpressional section of the San Andreas Fault is described by Meisling and Weldon

(1989). Low-angle thrust complexes on the Alpine Fault are mentioned by Wellman (1955) and Suggate (1963), but in these cases were ascribed to gravity collapse.

In this paper we present the results of detailed field mapping and structural investigation of a small, oblique thrust complex on the central Alpine Fault, and show both its relationship to transpressional motion and the influence of erosional and depositional processes on its structural evolution. The features described offer insights into the near-surface behaviour of major oblique-slip faults and the interaction of erosion and faulting. The study also highlights the range-front processes implied in the models of Koons (1990) and Beaumont *et al.* (1992), helps explain the regionally straight trace of the Alpine Fault and possibly why it has not partitioned into parallel strike-slip and thrust structures.

ALPINE FAULT TECTONICS

The Alpine Fault extends along the west coast of South Island and forms the major locus of displacement between the Australian and Pacific plates (Norris *et al.*, 1990; Berryman *et al.*, 1992). In the central section of the fault, the interplate vector calculated from the Nuvel-1 global model (DeMets *et al.*, 1990) trends 071°–251° with a velocity of 39 mm/year, corresponding to a right-lateral strike-slip component of 36 mm/year and a convergence component of 11 mm/year on the Alpine Fault. Not all of this occurs as slip on the Alpine Fault, as deformation is recorded over an area to the east at least 100 km wide (Walcott, 1978a; Norris *et al.*, 1990), but data from the fault trace (Berryman *et al.*, 1992; Cooper and Norris, 1994, 1995; Sutherland, 1994) suggest that 50–80% of interplate slip has been accommodated by the Alpine Fault since at least mid-Pliocene time. Repeated geodetic measurements across the central Alpine Fault allow calculation of shear strain rates in the range 0.7–1.4 μrad/year for various networks and periods, with an axis of maximum horizontal relative shortening oriented between 110° and 120° (Walcott, 1978a; Wood and Blick, 1986; Pearson, 1994).

Structural and geophysical data from the west coast suggest that the fault dips at angles of 45–60°SE through the upper crust (Sibson *et al.*, 1979; Woodward, 1979; Allis, 1986; Reyners, 1987; Norris *et al.*, 1990) and accommodates both strike-slip and convergent components (Norris *et al.*, 1990; Berryman *et al.*, 1992; Norris and Cooper, 1995). The Pacific plate is translated southwestwards parallel to the fault and thrust up and over the Australian plate, resulting in rapid uplift and exhumation of the hangingwall schists (Adams, 1981; Holm *et al.*, 1989; Tippett and Kamp, 1993) at rates of 5–10 mm/year (Bull and Cooper, 1986; Simpson *et al.*, 1994). The rapid uplift has resulted in the Southern Alps, a narrow mountain range with summit heights up to 3470 m, forming within the last 5 Ma. Schists of amphibolite facies have been exhumed from depths of 20–25 km during the last 2–3 Ma, and now outcrop as a narrow strip along the western side of the Southern Alps adjacent to the fault (Cooper, 1980; Holm *et al.*, 1989; Grapes and Watanabe, 1992). These schists forming the hangingwall of the fault are known as the Alpine Schists, and westward pass gradationally into mylonites within 1 km or so of the fault trace (Sibson *et al.*, 1979).

On a gross scale, for instance as seen from space, the trace of the Alpine Fault is remarkably straight. However, within the central part of the Southern Alps, in the vicinity of the Fox and Franz Josef glaciers (Fig. 1), the fault trace forms a series of en échelon segments on the scale of 1–5 km, with more northerly striking oblique thrust sections linked by more easterly striking strike-slip faults (Fig. 1) (Norris and Cooper, 1995). The slip direction on both types of segment is subparallel to the plate vector. Norris and Cooper (1995) suggest that this structure may be characteristic of transpressional faults, but that localization of the segments may relate to erosional topography across the fault trace.

Erosion on the western side of the Southern Alps is severe where rainfall exceeds 10 m/year (Griffiths and McSaveney, 1983). The high rainfall also results in dense forest growth below 1000 m, so that exposure below tree-line is mainly limited to rivers and streams. The Waikukupa River (Fig. 1) is deeply incised within the hangingwall schists and cuts across one of the oblique thrust segments of the fault trace on its way to the sea. Owing to dissection by small streams flowing into the Waikukupa River, the area around the fault trace is relatively well exposed.

Basement rocks west of the Alpine Fault are early Palaeozoic greywackes and their hornfelsed equivalents intruded by a series of intermediate to acid plutons (Western Province of Landis and Coombs, 1967). These are overlain by a series of fluvioglacial gravels belonging to the last glacial period (Otira glaciation *ca* 80–15 ka; Suggate, 1990) which may exceed 300 m in thickness (Warren, 1967). In the vicinity of the Waikukupa River, schist and mylonite in the hangingwall of the fault are thrust obliquely over these gravels and form imbricate thrust sheets which extend up to 2 km west of the main range front. Similar structures were first reported by Wellman (1955) and Suggate (1963). We have mapped these features in detail and investigated their structure and kinematic evolution.

FAULT ZONE STRUCTURE

Gross geometry of the fault zones

A distinctive sequence of fault rocks occurs within the hangingwall along oblique thrust segments of the Alpine Fault (e.g. Cooper and Norris, 1994). A basal cataclastic zone, from 2 to 40 m thick and hydrothermally altered to a grey or blue–green gouge, is overlain by fractured and faulted mylonite. The mylonite and ultramylonite is

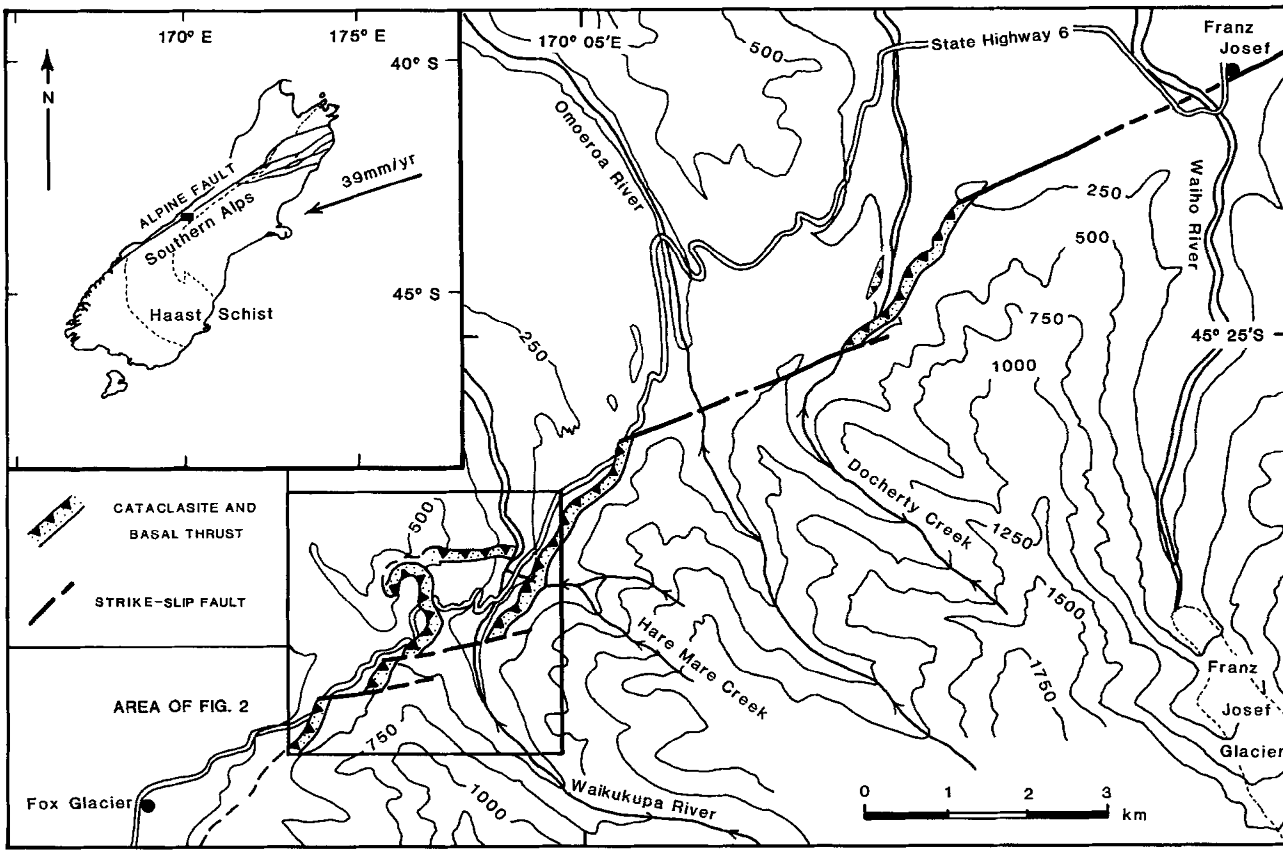

Fig. 1. Map of the Alpine Fault between Fox and Franz Josef glaciers, showing oblique thrust and strike-slip segments. The box outlines the area of Fig. 2. Inset map of South Island, New Zealand, shows the Alpine Fault, location of the main map (black rectangle) and the present plate vector (after DeMets *et al.*, 1990). Note that the centre line of the Waikukupa River valley is offset dextrally by approximately 1.5 km across the Alpine Fault zone.

derived from Alpine Schist and is commonly imbricated along gouge-filled shears for a distance of up to 100 m from the cataclasite. The degree of fracturing and imbrication diminishes up-section, and the mylonites become fairly uniformly dipping at a moderate angle to the southeast. Several hundred metres from the base, layers and lenses of weakly mylonitized schist appear within the mylonite, and the mylonite fabric is less intensely developed. A distinctive schist, with spaced shears separated by zones of sigmoidal schistosity (*S–C* mylonite), is known locally as 'curly schist' (Reed, 1964) and passes eastwards into normal, coarse-grained Alpine Schist. The whole mylonite zone may reach 1 km in thickness.

Within the Waikukupa River area, the Alpine Fault zone consists of two subparallel faults along which this sequence of cataclasite and mylonite is repeated (Fig. 2). The faults dip to the southeast at 25–30° and both emplace the mylonite–cataclasite sequence over Late Quaternary fluvioglacial gravel (Fig. 3). We term the western fault zone the Waikukupa thrust and the eastern zone, well exposed in Hare Mare Creek, the Hare Mare thrust. The Hare Mare thrust overlies the Waikukupa thrust and the sequence of fault rocks is repeated.

The Waikukupa thrust overrides fluvioglacial sediments consisting mainly of coarse, schist-derived gravels with rare sand and silt horizons (Okarito Formation in Fig. 2). Close to the thrust, the underlying gravels dip by as much as 50°W. The Hare Mare thrust also locally overrides fluvioglacial gravels of broadly similar character (Figs 2–4) which, however, rest unconformably on the mylonites of the Waikukupa thrust sheet at elevations well below its leading edge. Hence, they must post-date the latter's emplacement. These younger gravels (Moana Formation in Fig. 2) are also deformed and tilted to near-vertical immediately below the Hare Mare thrust (Fig. 5). The Hare Mare thrust is consequently the younger structure.

Measurements of exposed basal thrust surfaces of both thrust sheets show an average dip of 25–30°SE (Fig. 6a). Excluded are measurements of the Waikukupa thrust at its western extremity, where it overrides the original land surface and becomes low dipping and of variable strike. The few measurements of well-preserved striations within the basal gouge indicate a transport direction of the thrusts towards 250–260° (Fig. 6a). This is oblique to the general dip direction of the thrust planes, but is broadly consistent with the direction of predicted plate convergence (DeMets *et al.*, 1990).

Figure 7 is a structure contour map on the Waikukupa thrust and western part of the Hare Mare thrust constructed from surface exposure. Figure 8 is an attempt to show the three-dimensional geometry of the fault system utilizing the data in Fig. 7. At its western

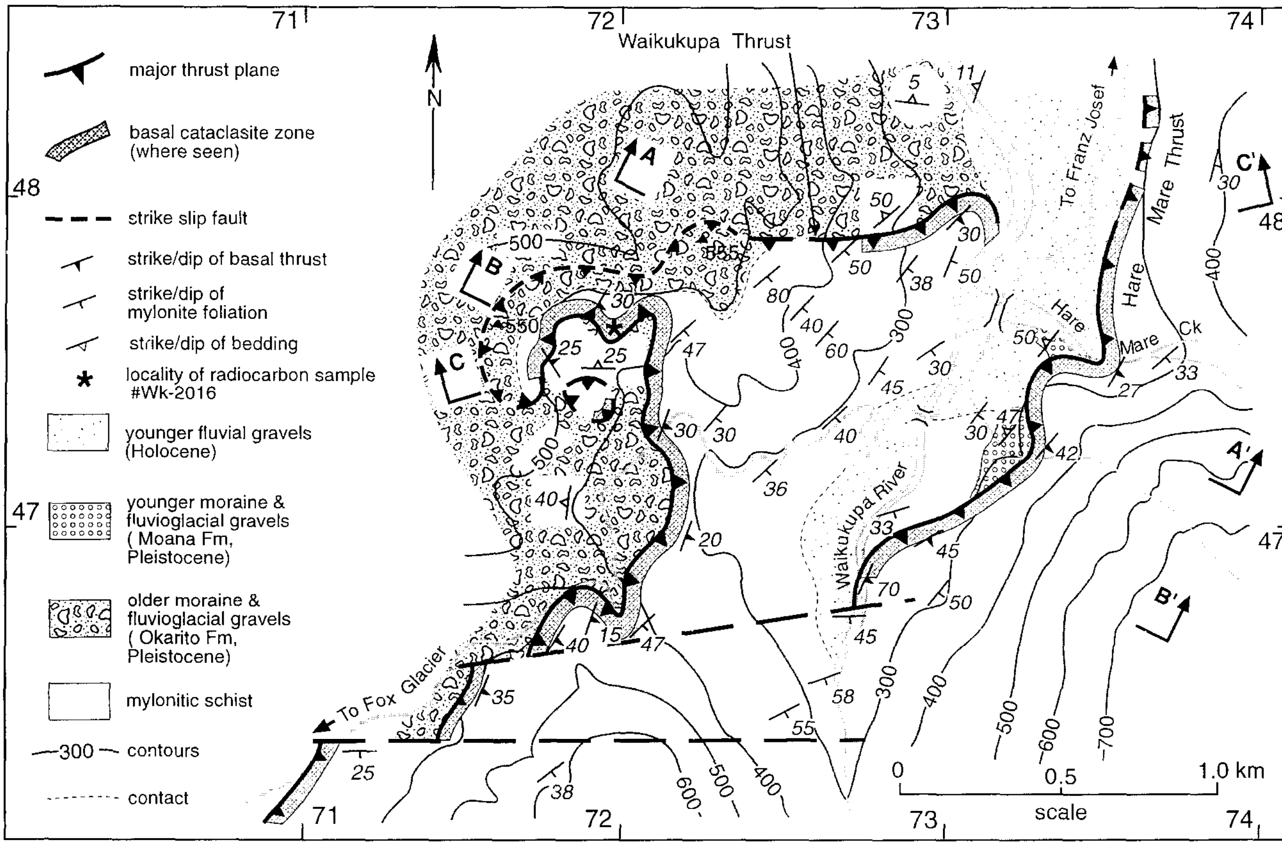

Fig. 2. Geological map of the Waikukupa River area (see Fig. 1 for location). The western thrust zone is the Waikukupa thrust and the eastern zone the Hare Mare thrust. Note that the basal cataclasite is not shown to scale—it varies from 2 to 10 m in thickness. Numbered ticks on the margin are 1000-m gridlines of the New Zealand national metric grid.

extremity, the Waikukupa thrust becomes almost horizontal as it reaches the original land surface on top of the Okarito Formation moraines. The Hare Mare thrust is a younger imbricate of the Waikukupa thrust and presumably joins it at depth. It is a prominent feature as far south as the Waikukupa River. Here, it steepens to near vertical and is abruptly truncated by an E–W-striking, steeply dipping fault zone (Fig. 2). The mylonitic foliation within the Waikukupa thrust sheet is locally rotated to steep attitudes within a few tens of metres of the fault zone. Farther west, the zone of faulting is visible as a linear topographic feature from the air. Where this zone crosses the Waikukupa River, a prominent subvertical fault approximately 2 m wide is exposed, containing crush material and clay gouge (Fig. 9). This fault zone strikes on average about 080° and dips 65–70°N. Slickenside striations within the gouge plunge about 20° towards 070° (Fig. 6b), very similar to striations on the basal thrust (Fig. 6a).

Southwest of the E–W fault, there appears to be only one thrust zone more or less continuous with the Waikukupa thrust and coincident with the range front. The Hare Mare thrust is coincident with the present range front to the north of the E–W fault, and here the Waikukupa thrust is highly dissected and apparently inactive. The E–W strike-slip fault zone has therefore the form of a tear within the Waikukupa thrust sheet, transferring motion from the Hare Mare thrust onto the southern part of the Waikukupa thrust. In classic thrust geometry terms, it could be described as a steep lateral ramp (Boyer and Elliott, 1982). As noted above, slip directions on the thrusts and on the linking strike-slip fault are subparallel, resulting in a kinematically stable system of non-parallel faults.

Internal geometry of the Waikukupa thrust sheet

The mylonites comprising the Waikukupa thrust sheet are cut by large numbers of gouge-bearing faults and folded into angular folds on a scale of 10 cm–20 m (Fig. 10). The dominant faults show reverse separation and are associated with the limbs of angular folds. Whereas the mylonitic foliation in general dips at 20–30°SE, subparallel to the basal thrust, it has been systematically rotated to steeper angles and folded adjacent to reverse faults, forming a series of fold–thrust packets. This structural style is particularly well developed in that part of the thrust sheet north of the E–W fault.

Well-exposed sections through the lower part of the thrust sheet on the west side of the Waikukupa River

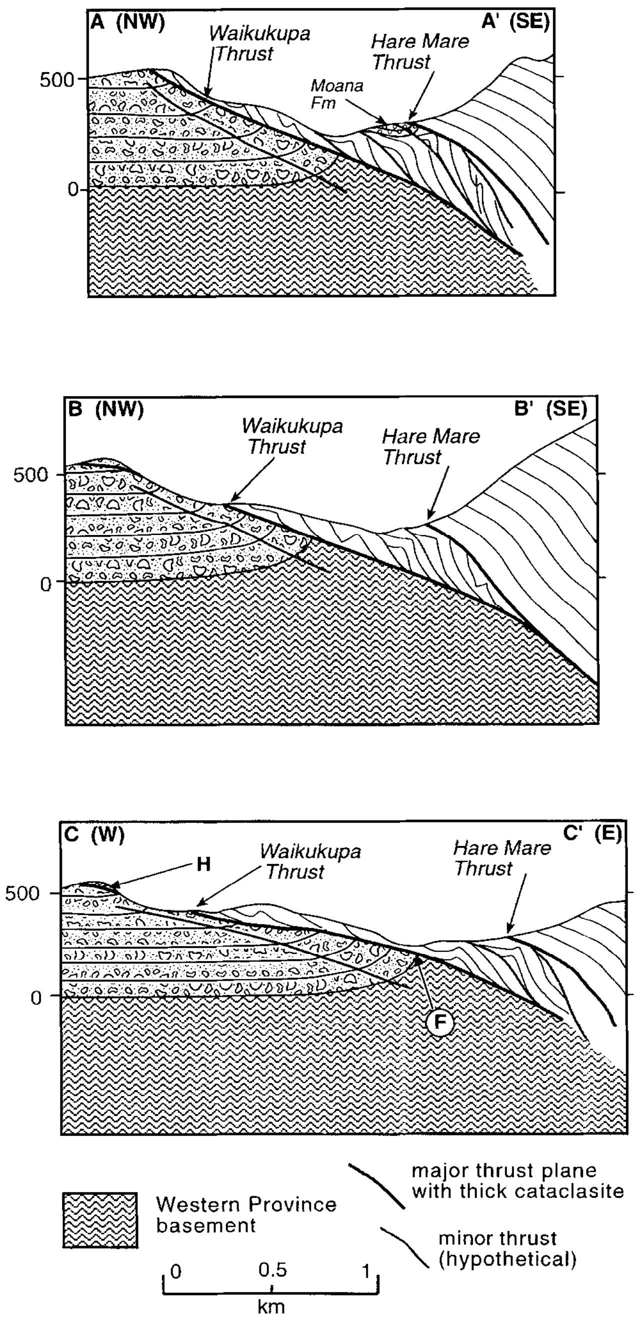

Fig. 3. Cross-sections along the lines of section shown in Fig. 2. Note that only section C–C′ is approximately parallel to the direction of transport of the thrusts. The location of the base of the Okarito Formation is somewhat conjectural, as it does not outcrop in the area of the map. Secondary thrusts shown within the gravel units are partly conjectural, although minor faults are seen locally within the tilted gravel units (e.g. Figs 4 & 6). Ornament the same as in Fig. 2 except for Western Province basement.

were recorded onto a mosaic of photographs to construct a composite structural section illustrating the general style of deformation (Fig. 11). The geometry has elements of a duplex structure, although ramp faults do not always extend from the base but may nucleate within a fold. Shortening of the well-foliated mylonite has led to buckling instabilities, in turn nucleating brittle faults. From offset of marker horizons such as mylonitized pegmatite veins, the displacement on these faults does not appear to have been more than a few metres. We estimate shortening due to internal deformation within this part of the Waikukupa thrust sheet to be 15–20% in cross-sections perpendicular to the fold hinges

Poles to the foliation lie on a well-defined great circle with an axis of folding of 10°/049° (Fig. 12a), roughly coincident with measured fold hinges. Measurements of quartz stretching lineations, unevenly developed within the mylonites, spread along a small circle due to rotation with the foliation. Most of the foliation data come from the area north of the road (Fig. 2) so that the axis of folding is approximately parallel to the strike of the basal thrust here.

Poles to reverse faults also define a great circle with a broadly similar orientation to that of the foliation (Fig. 12b). Most faults dip SE more steeply than the average foliation dip, with a small group of antithetic faults dipping NW. Slickenside striations on reverse separation faults are spread between orientations similar to those on the basal thrust and orientations perpendicular to the fold axis.

A large number of steep faults lacking evidence of reverse separation also cut the mylonites. They have a preferred strike between NE and SE, with modes striking around 140° and 060° (Fig. 12c). Slickenside lineations occur rarely within gouge and indicate a dominant strike-slip character. The few offset observations suggest a component of dextral slip on the more NE- and E-striking faults, and of sinistral slip on the SE-striking faults. Together with the reverse faults, they are consistent with an ESE shortening.

In the vicinity of the E–W strike-slip zone along the Waikukupa River, the foliation becomes rotated and folded to steep attitudes along numerous secondary Riedel shears adjacent to the main fault zone. The sense of rotation is consistent with dextral shear within the zone (Fig. 12d).

Internal geometry of the Hare Mare thrust sheet

The gravels below the thrust are tilted and exhibit minor thrusts subparallel to the main thrust plane (Figs 4 & 5). In thin micaceous silt horizons, a spaced crenulation cleavage structure is locally developed due to internal shearing and grain rotation (Fig. 13a & b). The orientation of the cleavage is compatible with shear parallel to the striations on the basal fault gouge (Fig. 14a). Bedding in the gravels is tilted about an axis which is similar to the fold axis in the underlying Waikukupa thrust sheet and oblique to the cleavage–fault intersection, although it does lie approximately within the plane of cleavage.

Poles to bedding in both sets of deformed gravels (underlying the Waikukupa and Hare Mare thrusts) show a fairly wide scatter although generally dipping to the northwest (Fig. 14b). The Okarito Formation beneath the Waikukupa thrust generally has a more

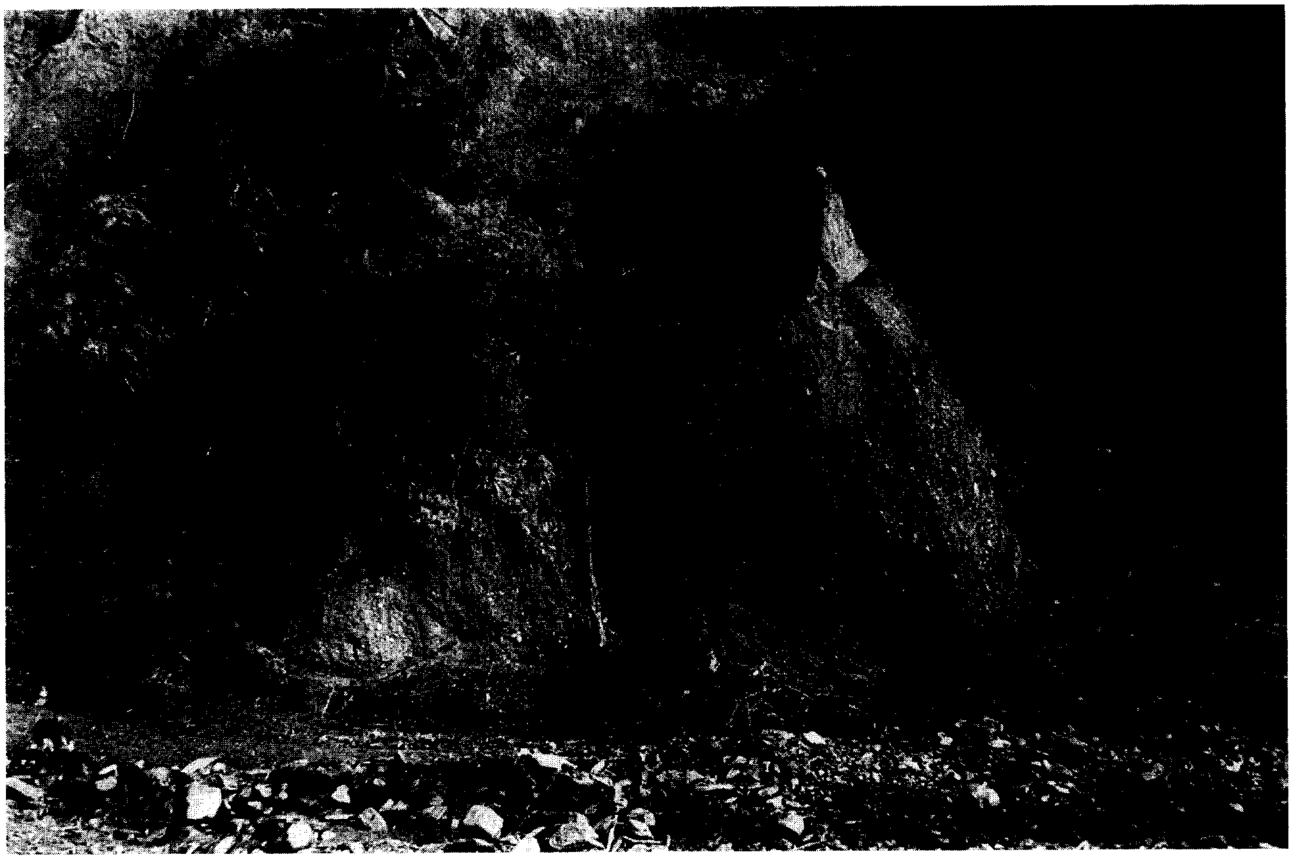

Fig. 4. View, looking southwest, of outcrop of the Hare Mare thrust (arrowed) in Hare Mare Creek. The light-coloured band directly above the thrust is basal cataclasite emplaced over steeply dipping gravels. Note the minor thrust plane within gravels. The outcrop is approximately 10 m high. Figure 5 is a structural section through this outcrop.

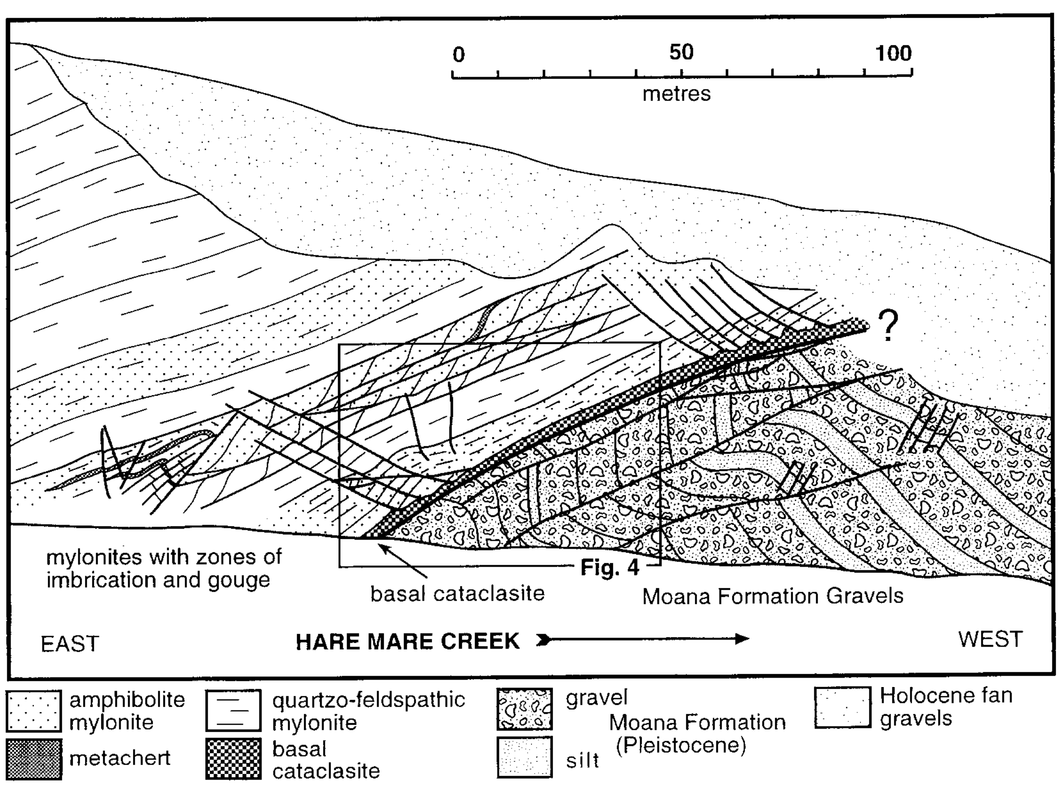

Fig. 5. Geological cross-section parallel to Hare Mare Creek (see Fig. 4) based on field observations and constructed from photographs corrected to a vertical plane by radial-line plotting. The view is to the southwest. Note that the ornament differs from Fig. 2.

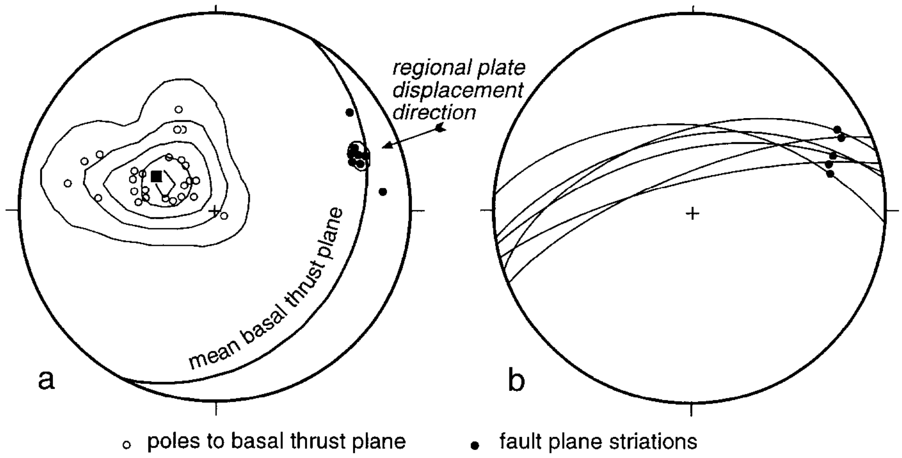

Fig. 6. (a) Equal-area plot of poles to the basal thrust plane (open circles, 25 measurements), and slickenside striations (filled circles, 8 measurements) within the basal gouge for both the Waikukupa and the Hare Mare thrusts. Contours for the poles to thrust planes are in intervals of 3 × uniform. Mean thrust plane (great circle, pole = filled square), 030°/28°SE; mean striation, 22°/070°. (b) Equal-area plot of gouge zones within strike-slip fault zone in the Waikukupa River, together with slickenside striations (filled circles).

northerly strike than the Moana Formation beneath the Hare Mare thrust. The bedding may have had a significant and variable depositional dip in the vicinity of the fault.

Relatively undeformed gravels, mainly fan gravels, locally overlie the Hare Mare thrust (Fig. 5). These also unconformably overlie the deformed Moana Formation sediments. Unfortunately their contact with the fault plane itself is obscured but, in the Hare Mare Creek section (Fig. 5), their base appears displaced by about 20 m vertically across the trace of the fault. While they clearly were formed after the Moana glacial retreat, we have been unable to date them any more precisely.

RATES OF DISPLACEMENT

The ages of the gravels overthrust by mylonite along the two thrust surfaces are uncertain. The gravels beneath the Waikukupa thrust are derived from low- to medium-

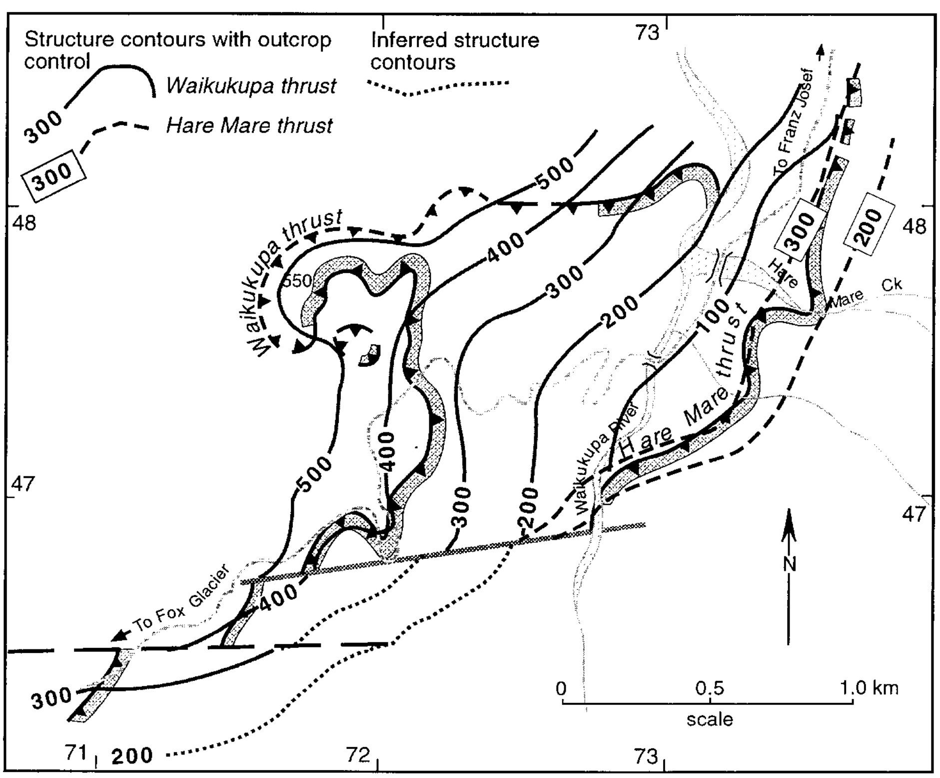

Fig. 7. Structure contour map of the Waikukupa and Hare Mare thrusts; the same area as shown Fig. 2. Contours are reasonably well constrained by outcrop data except where shown as dotted lines. Numbered ticks on the margin are 1000-m gridlines of the New Zealand national metric grid.

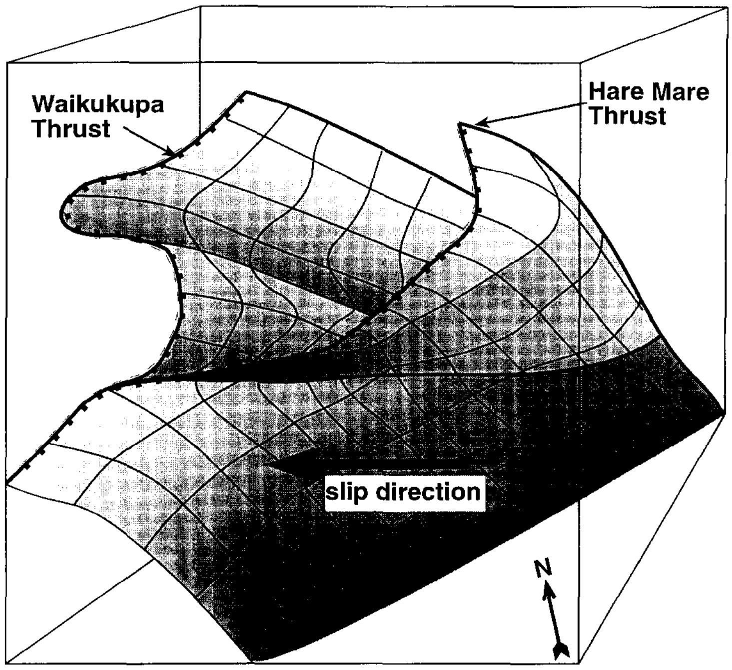

Fig. 8. Three-dimensional reconstruction of the thrust geometry based on the structure contours in Fig. 7. The diagram is drawn looking north. Grid lines parallel to the strike and dip of the thrust surface are purely for diagrammatic effect and have no other significance.

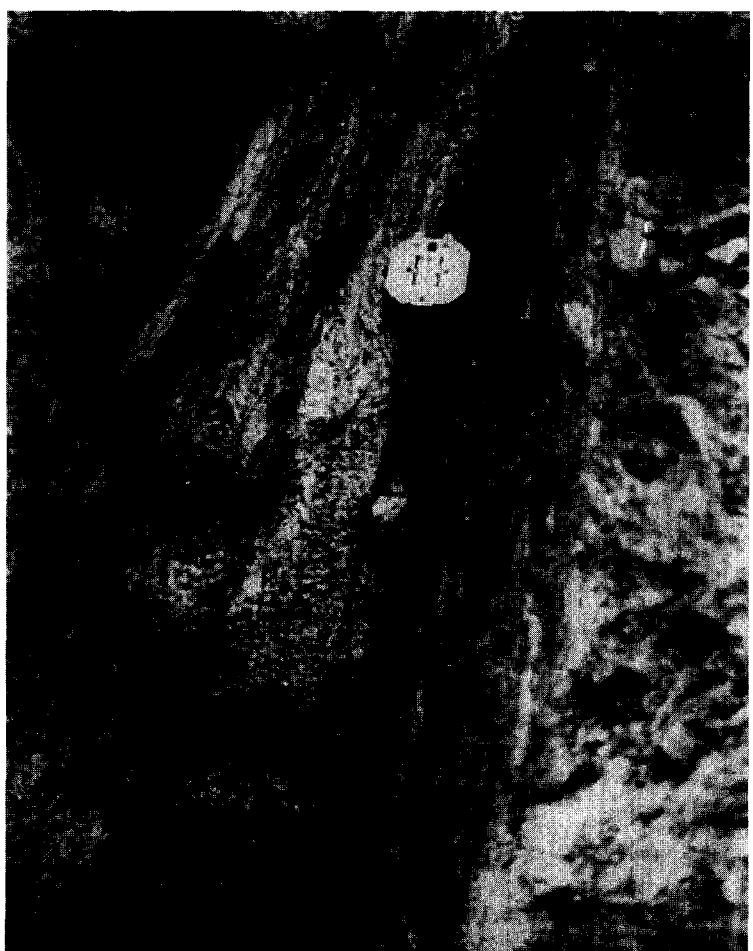

Fig. 9. Subvertical fault plane with gouge which strikes approximately E–W and coincides with the disappearance of the Hare Mare thrust in the Waikukupa River (see Fig. 2).

Fig. 10. Large angular fold and minor thrusts within imbricated mylonites of the Waikukupa thrust sheet. View to the southwest, 1.75-m tall person for scale (see Fig. 11 for a structural section).

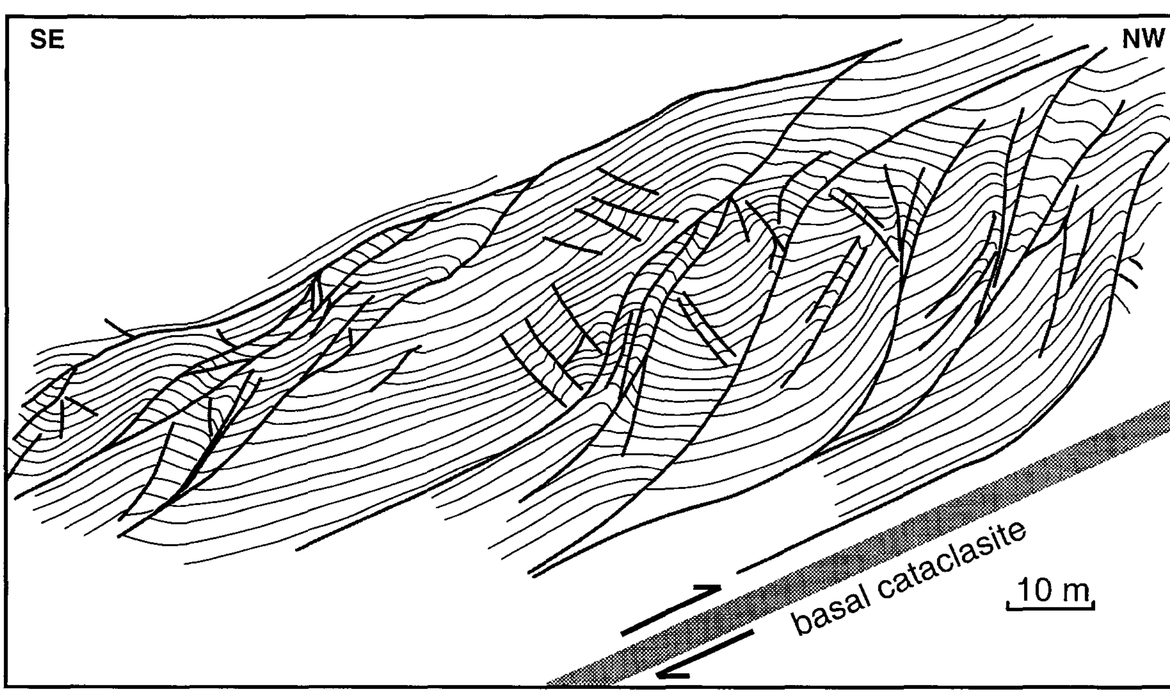

Fig. 11. Typical cross-section through mylonites near the base of the Waikukupa thrust showing angular folding and duplex-like geometry. The section is constructed from field-annotated photographs assembled into a single section (e.g. Fig. 10). Thick lines are faults, commonly containing a thin (1–5 cm) gouge layer; thin lines are form lines parallel to the mylonitic foliation.

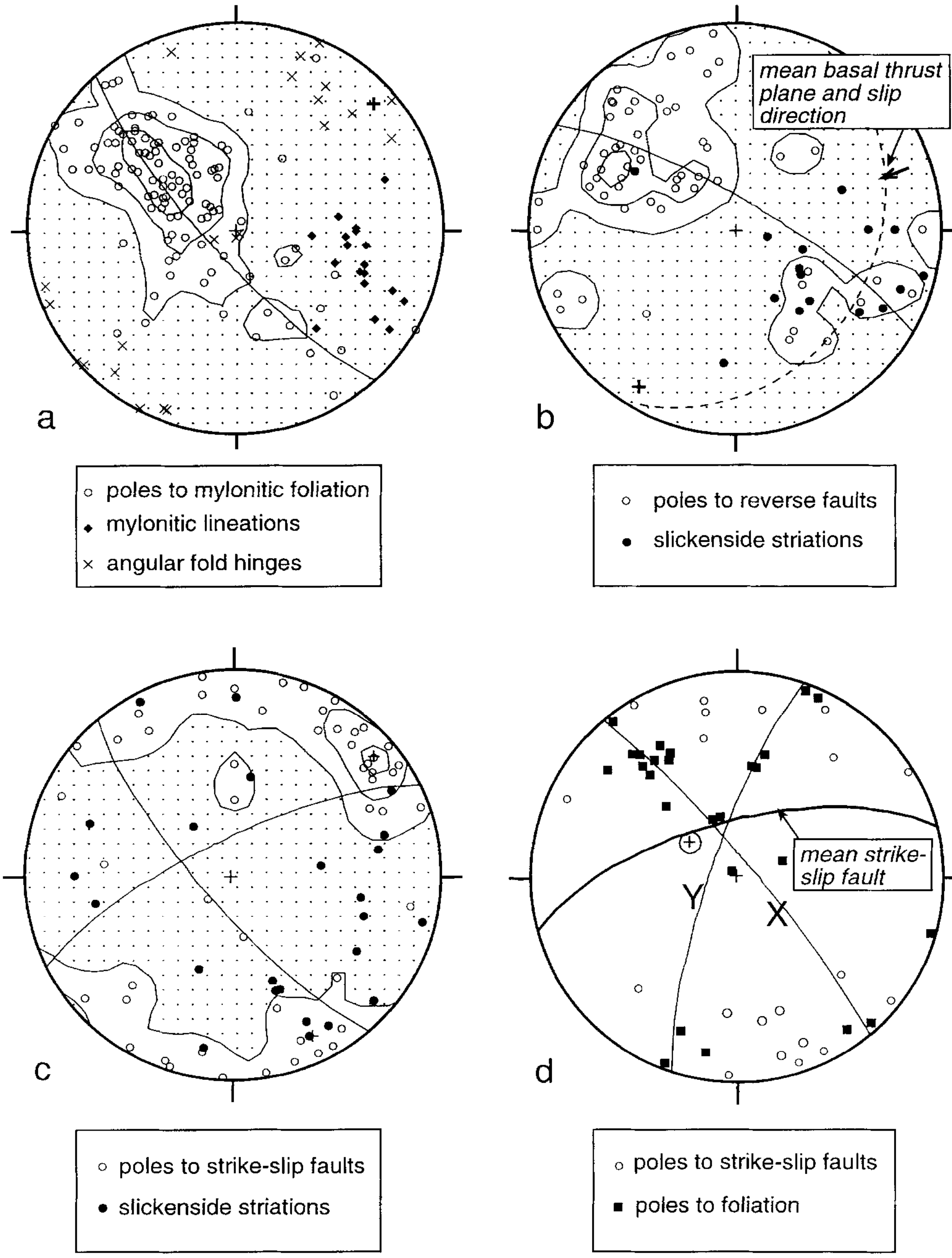

Fig. 12. Equal-area plots of data from the Waikukupa thrust sheet. (a) Poles to mylonitic foliation (open circles, 97 measurements) and hinges of angular folds (crosses, 23 measurements). Contours around foliation poles are in intervals of 3 x uniform. Best-fit girdle through poles has an axis (bold cross) of 10°/048°. Mylonitic stretching lineations (filled diamonds, 19 measurements) lie on a crude partial small circle around the fold axis. (b) Poles to reverse-separation faults (open circles, 51 measurements) together with related slickenside striations (filled circles). Principal modal plane, 030°/56°SE; best-fit girdle through poles has an axis (bold cross) of 11°/210°. (c) Poles to non-reverse faults (open circles, 53 measurements) in the Waikukupa thrust sheet, together with slickenside striations (filled circles, 25 measurements). Two main modal planes are shown as great circles. (d) Poles to foliation (solid squares) and steep fault planes (open circles) in subarea close to the E–W fault. The bold great circle is the mean orientation of the strike-slip zone. Encircled cross is the modal orientation of foliation within the whole thrust sheet from (a). Foliation has been rotated to steep dips mainly about a NE–SW axis (girdle X), consistent with right-lateral shear. Minor girdle (Y) represents the local rotation of foliation towards parallelism with secondary (?Riedel) shears within the array.

grade schist typical of that exposed in the range front of the Southern Alps. They do not contain mylonitic debris and hence were not derived directly from the fault zone. They contain sandy horizons and are at least partly fluvial in origin. Other parts contain large boulders, are poorly sorted and are probably moraine.

Two major suites of fluvioglacial gravels are mapped regionally in the area (Warren, 1967), corresponding to the two major advances during the last (Otira) glaciation. The main ridges extending westwards from the Alpine Fault to the sea are formed from the older Okarito Formation, whereas the younger Moana Formation

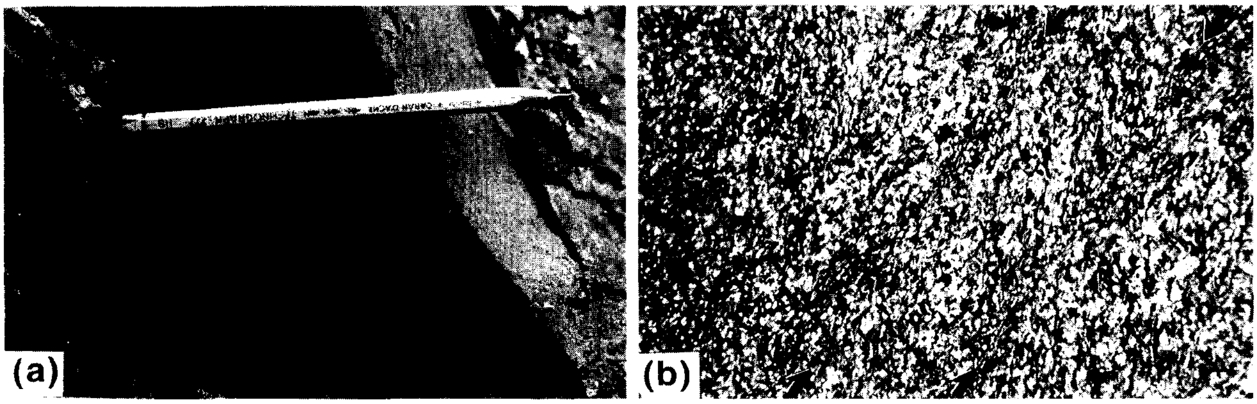

Fig. 13. (a) Crenulation cleavage structure (subvertical) within a micaceous sandy silt horizon below the Hare Mare thrust (see Fig. 14). (b) Thin section of micaceous silt, showing crenulation cleavage structures. Cleavage planes extend diagonally from top right to bottom left (examples arrowed). Plane-polarized light; the base of the photograph is 10 mm across.

occurs extensively along valley sides (Warren, 1967). The ages of the two formations depend on correlations with dated sequences elsewhere and on the chronology of late Pleistocene glaciations (Suggate, 1990). The best estimates are 65–75 ka for the older advance, correlated with O-isotope stage 4, and 14–18 ka for the younger advance (Suggate, 1990). Deposits from older glacial periods are not recognized in central Westland, and have presumably been removed during the ice advance of the Otira glaciation.

A piece of wood embedded within the basal gouge of the Waikukupa thrust in contact with the underlying gravel (Fig. 2; University of Waikato Radiocarbon Laboratory number Wk-2016) gave a ^{14}C age of 40,000 years BP, at the upper limit of reliable radiocarbon dates. Because of the possible effect of contamination, this date must be viewed with caution. It does indicate, however, that the overthrust gravels are older than 40 ka, and hence most likely to belong to the Okarito Formation.

Although the gravels beneath the Hare Mare thrust are very similar in general character to those described above, they rest unconformably, almost at present river level, on the mylonites of the Waikukupa thrust sheet, and hence must be younger. Their relative age and position within the valley suggests correlation with the

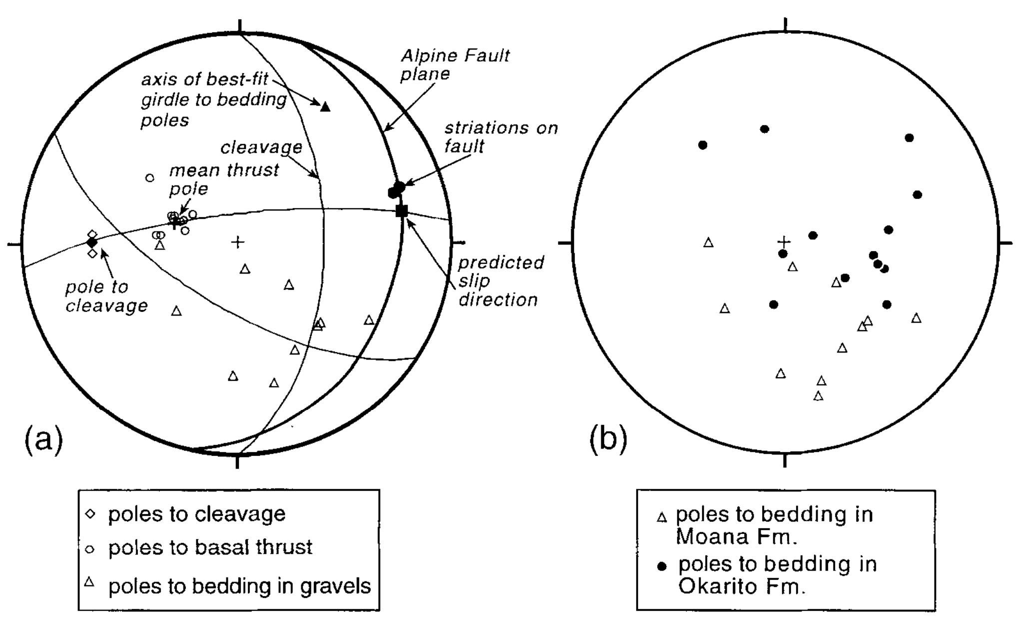

Fig. 14. (a) Equal-area plot of structural features of the Hare Mare thrust zone exposed in Hare Mare Creek. Open circles/cross, poles to basal thrust plane/mean pole; filled circles, slickenside striations on basal thrust; open diamonds/filled diamond, poles to crenulation cleavage structure in footwall silt horizon/mean pole (see Fig. 13); open triangles, poles to bedding in underlying Moana Formation gravels. The solid square represents the inferred shear direction in the thrust zone perpendicular to the intersection of the cleavage and thrust plane (assuming a simple-shear model of deformation). (b) Equal-area plot of poles to bedding in fluvioglacial gravels. Open triangles, Moana Formation beneath the Hare Mare thrust; closed circles, Okarito Formation beneath the Waikukupa thrust.

Moana Formation. Unfortunately no datable material has yet been found within them.

Figure 3(c), a cross-section drawn parallel to the slip vector on the basal thrust plane (Fig. 6a), shows the location of the footwall cutoff point (F in Fig. 3c) of the underlying gravels approximately at river level—this is a conservative estimate as it could be deeper. The hangingwall cutoff point (H in Fig. 3c) makes the assumption that gravel remnants overlying mylonite at the leading edge of the Waikukupa thrust (Fig. 2) are the displaced equivalents of those beneath. The distance between points F and H measured parallel to the slip vector is 1400 m. If this displacement is considered to have occurred prior to the Moana glaciation (i.e. between 70 and 15 ka), then a minimum slip rate of about 26 mm/year is indicated.

The minimum amount of displacement on the Hare Mare thrust over the Moana gravels is about 150 m. Combining the minimum displacements on both thrusts gives a minimum displacement rate of approximately 22 mm/year (± 2 mm/year depending on the precise age of the Okarito Formation). This is very much a minimum, however, and does account for the amount of displacement on the Hare Mare thrust prior to the basal cataclasite reaching the surface, nor for any shortening within the Waikukupa thrust sheet as evidenced by the extensive folding and thrusting. Shortening estimates within sections of the Waikukupa thrust sheet (e.g. Fig. 11) are 15–20%. If this is maintained over 1400 m, a further 250 m of displacement is indicated. The total displacement on the Hare Mare thrust is difficult to determine but would appear from cross-sections (Fig. 3) to be at least 500 m. These additional displacements would increase the slip rate to around 30 mm/year. A slip rate of 22–30 mm/year therefore can be regarded as a reasonable estimate despite the uncertainties in the data (the probable age range of the Okarito Formation gives an additional uncertainty of ± 2 mm/year to these figures). This is comparable to displacement rates determined elsewhere in central Westland (Cooper and Norris, 1994, 1995). These rates represent between 60 and 75% of the calculated rate of plate displacement (Nuvel-1 solution, DeMets *et al.*, 1990).

The Waikukupa River channel is also displaced dextrally across the fault zone (Fig. 1) by about 1.5 km. As the valley is excavated in the Okarito Formation deposits, the displacement must post-date this glacial advance and hence has occurred during the past 50–60 ka, commensurate with the rates calculated above.

STRUCTURAL EVOLUTION

Emplacement of the Waikukupa thrust

The continuity of the near-surface thrust plane with the main Alpine Fault at depth is indicated by the sequence of fault rocks emplaced along it. The mylonites within the hangingwall are ductily deformed rocks with amphibolite-facies assemblages (Sibson *et al.*, 1979; Cooper and Norris, 1994; Grapes and Watanabe, 1994) and are likely to have formed at depths greater than 8 km (Koons, 1987; Holm *et al.*, 1989). The hydrothermally altered green cataclasite consists of greenschist-facies assemblages (Grapes, 1993; authors' unpublished observations). Its thickness and alteration suggest formation by cataclasis and retrograde metamorphism at depths of several kilometres, even after allowing for elevated isotherms (Koons, 1987; Holm *et al.*, 1989) and circulating hot fluids (Johnstone *et al.*, 1990). It is likely therefore that the typical hangingwall sequence of fault rocks is mainly formed at depths much greater than 1 km, with near-surface deformation producing clay gouge on major thrust planes and the development of secondary gouge-filled fractures.

During emplacement of the Waikukupa thrust sheet, some deformation of the footwall occurred as the gravels have been tilted westward. The present top of the Okarito moraine west of the thrust is thought to represent the original land surface. Presumably gravels uplifted in the hangingwall were largely removed by erosion, so that eventually the leading edge of the mylonites and cataclasites reached the original land surface where the thrust flattens out. It is here that locally we find deformed gravels overlying the mylonites in the hangingwall, which we interpret as remnants of the uplifted gravels. The occurrence of these gravels at the present western edge of the thrust where it flattens over the footwall moraine surface suggests that the emplacement of the thrust sheet ceased when the mylonites reached the land surface.

Erosional control on imbrication of the Waikukupa thrust

The abandoned portion of the Waikukupa thrust, and the development of the Hare Mare thrust, coincide with the area cut through by the Waikukupa River valley. The Waikukupa River is offset dextrally between the two thrusts by an amount equivalent to the larger part of the post-Okarito displacement. Thus, the river was actively maintaining its valley, at least across the hangingwall, for most of this time. The presence of the younger Moana Formation gravels at low levels within this valley indicates that, during the Moana glacial advance, the valley was cut to within 50 m of its present level. Active erosion of the thrust sheet was therefore occurring during its emplacement, and at an average rate greater than the uplift rate on the thrust.

By the time of the Moana glaciation, erosional downcutting by the Waikukupa River had reduced the thickness of the thrust sheet to the extent that it became subcritical (Davis *et al.*, 1983; Dahlen, 1984) and could no longer behave as a structurally coherent entity. We relate the development of the prominent folding, faulting and imbrication within the abandoned portion of the Waikukupa thrust sheet to loss of mechanical integrity due to erosion by the Waikukupa River. This process eventually resulted in the abandonment of the Waiku-

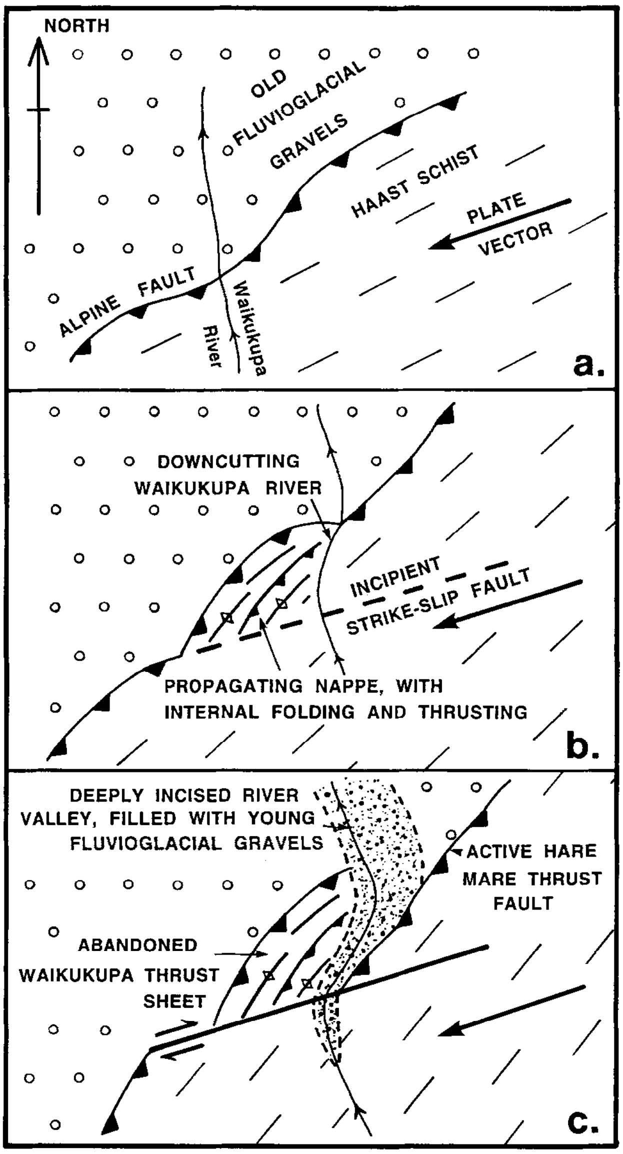

Fig. 15. Interpreted evolution of the Waikukupa section of the Alpine Fault. (a) Emplacement of the Waikukupa thrust sheet following deposition of the Okarito Formation. (b) Downcutting by the Waikukupa River leads to shortening and imbrication within the Waikukupa thrust sheet. (c) Emplacement of the Hare Mare thrust sheet over Moana gravels and the abandonment of the northern part of the Waikukupa thrust.

kupa thrust and the development of a major new imbricate thrust, the Hare Mare thrust (Fig. 15b & c). This conclusion is consistent with the folding and tilting within the Moana Formation gravels beneath the Hare Mare thrust which must be coeval with deformation in the underlying mylonites. We also note that fold–thrust structures of this nature and intensity are only seen along the Alpine Fault in abandoned portions of thrust sheets. Active thrusts, such as the Hare Mare thrust and the Alpine fault at Gaunt Creek (Cooper and Norris, 1994), contain zones of internally imbricated foliation but not the large-scale folding and thrusting seen here. The active segment of the Waikukupa thrust south of the E–W shear zone also fails to display these features.

The Hare Mare thrust must root back onto the main fault surface as it emplaces the full sequence of fault rocks which, as discussed earlier, were developed on the Alpine Fault at depth. The development of the Hare Mare imbricate as the main zone of translation and the abandonment of the Waikukupa thrust must result from the failure of internal shortening within the latter to maintain criticality in the face of rapid erosion. The Hare Mare thrust thus represents an 'out of sequence' imbricate of the main fault zone, caused by erosional degradation of the leading thrust sheet (cf. Morley, 1988). We interpret the position of the Hare Mare thrust to be due entirely to the location of maximum erosion along the river valley. Continued movement on this thrust, rather than further propagation westward, is a result of continued removal of material at the range front. The mechanics are essentially the same as those invoked for the whole mountain range (Koons, 1990; Beaumont *et al.*, 1992), but on a much smaller scale.

Application of critical wedge models

Dahlen (1984) has presented an exact solution for a two-dimensional non-cohesive critical wedge relating the values of surface slope and basal detachment slope to internal and basal friction coefficients for wedges at equilibrium. Using this relationship, he has constructed a plot of basal against surface slope showing fields of stability for an internal friction angle of 30° and for various basal friction coefficients (Dahlen, 1984, fig. 11). For the purpose of demonstration, we have recalculated this diagram for an internal friction angle of 30° and basal friction angles of 15° and 20° (Fig. 16). Although the Waikukupa thrust sheet is strictly three-dimensional, the Dahlen treatment is a useful approximation (cf. Koons, 1994).

The average dip of the basal surface of the Waikukupa thrust obtained from the structure contour map (Fig. 7) is 25°. An estimate of the average surface slope of the wedge at the time of maximum excavation during the Moana advance is obtained from the topographic contours (Fig. 2) by measuring the gradient from the thrust tip to the base of the Moana Formation; this is approximately −14° (the negative value indicating a similar direction of slope as the basal thrust). For a basal friction angle of 10° and greater, a wedge of this shape plots within the unstable field of Dahlen (1984), and is predicted to deform by internal thrusting (Fig. 16). While we have no direct measurements of the friction coefficient of the basal gouge, values of friction angle for clay gouge on the San Andreas Fault measured by Morrow *et al.* (1982) range from 15° to 30°; gouge on the Alpine Fault is likely to be comparable. Fluid pressures are unlikely to be high on the base of a near-surface wedge. Allowing for the uncertainties in friction, and for the simplifying assumptions of a dry non-oblique wedge, the comparison nevertheless is consistent with the geological evidence,

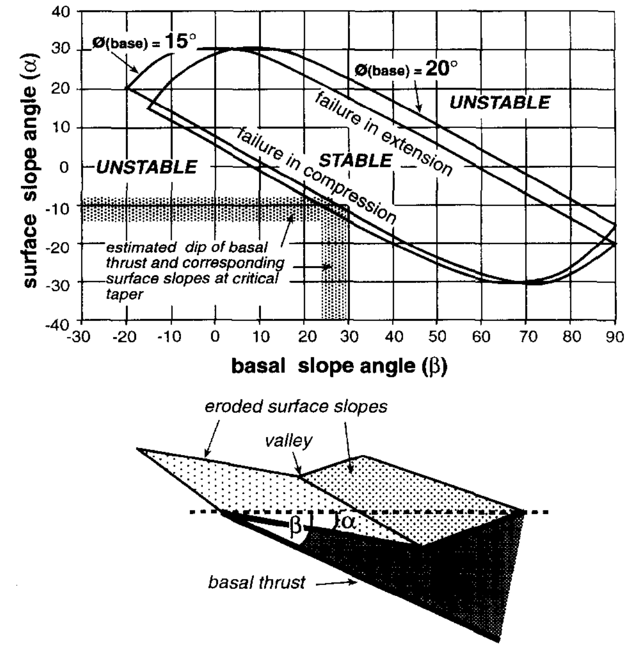

Fig. 16. Plot of surface (i.e. topographic) slope angle (α) against basal thrust dip (β) for critical wedges with an angle of friction internally of 30° and for angles of friction on the base of 15° and 20° (from Dahlen, 1984, equations 17–19, fig. 11). A positive value for α represents a topographic slope towards the apex of the wedge, a negative value away from this. The mean dip of the basal Waikukupa thrust and corresponding surface slope angles for criticality are highlighted. The lower diagram illustrates the idealized geometry of the Waikukupa thrust wedge in terms of the angles plotted. Note α for this case is negative. See text for discussion.

that erosion during the Moana advance rendered the thrust sheet subcritical.

With a basal fault dip of 25°, surface slopes of −8° and −11° are required for criticality for basal friction angles of 20° and 15°, respectively (Fig. 16). These would correspond to depths of erosion of the river below the old moraine surface of *ca* 100 and 150 m, compared with 250 m during the Moana advance and 300 m today. Hence, loss of criticality may have occurred prior to maximum excavation and resulted in progressive internal imbrication of the thrust sheet as the river eroded its valley laterally as well as vertically.

At its southern margin, the Hare Mare thrust terminates along the E–W strike-slip zone. The Alpine Schist within the uplifted hangingwall of the Alpine Fault contain numerous ENE-striking strike-slip faults (Hanson *et al.*, 1990; Norris and Cooper, 1995) and it is probable that the lateral shear zone was localized by one of these. Nevertheless, we argue that the termination of the Hare Mare thrust at the Waikukupa River is linked to the reduction in erosion rates southwest of the river valley where the Waikukupa thrust has maintained its mechanical integrity. The surface slope of the Waikukupa thrust wedge immediately north of the strike-slip zone is approximately −9°, whereas south of the zone it flattens rapidly to less than −5°. For reasonable friction coefficients, these slopes are consistent with a change from unstable to stable on the Dahlen model (Fig. 16).

Internal deformation of the thrust sheets

We interpret the internal deformation in the thrust sheets mainly to represent internal shortening during loss of criticality of the thrust wedge. In the case of the Hare Mare thrust, this may relate to strength variations on the base during propagation of the thrust surface, whereas the widespread internal deformation of the Waikukupa thrust sheet relates to loss of criticality due to erosional thinning (e.g. Davis *et al.*, 1983; Dahlen, 1984; Willemin, 1984; see above). During much of the thrust emplacement, however, we suggest that the thrust wedges moved as single entities, as predicted for oblique critical wedges by Platt (1993) and Koons (1994), with oblique-slip on the basal plane. Hence, large rates of displacement have been accommodated on a relatively simple and narrow structure.

Folding within the Waikukupa thrust sheet occurred close to the surface, in accordance with the model of Jamison (1992) for imbrication of layered sequences at shallow depths where low confining pressure allows easy slip along closely spaced layers. Foliation within the Hare Mare thrust has been imbricated within narrow discrete zones but not folded into large open-fold structures. This may reflect formation at greater depth with less direct influence of the free ground surface.

The amount of shortening within the Waikukupa thrust sheet as a result of folding and fault imbrication was estimated to be 15–20%. If this is distributed over the area of the thrust, it represents around 250 m of displacement. At a 25 mm/year displacement rate (see above), the structures could have formed in a minimum of *ca* 10 ka. There is no evidence along the Alpine Fault for fault creep. Instead, the fault appears to move during large, infrequent earthquakes (Berryman *et al.*, 1992). If this is the case here, then nucleation of the fold–fault structures must have occurred coseismically. The slickenside striations measured are mainly groove striations, rather than fibre lineations developed during creep deformation (e.g. Elliott, 1976).

Transpressional strain model for the Waikukupa thrust sheet

The extensive folds and thrusts within the Waikukupa thrust sheet are not oriented perpendicular to the basal slip direction. As the slip vector on the basal thrust is highly oblique to strike, one approach is to consider deformation of the stationary wedge as a distributed transpressional strain developed during periods when slip on the basal thrust was difficult due to a subcritical taper.

Fossen and Tikoff (1993) and Tikoff and Teyssier (1994) have modelled deformation in transpressional zones. Their models are based on homogeneous strain and involve a rectangular block with strain axes normal and parallel to the base. Accepting these simplifications for the moment, we consider a block bounded below by the basal thrust undergoing transpressional deformation

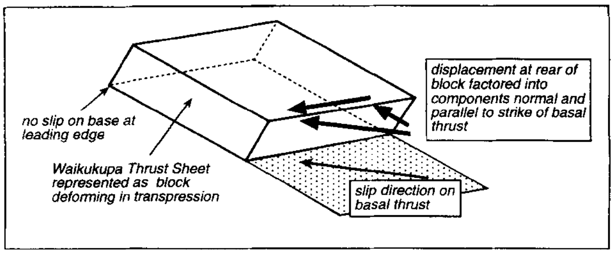

Fig. 17. Diagram showing the Waikukupa thrust sheet as a simplified block undergoing transpression (cf. Fossen and Tikoff, 1993). See text for discussion.

represented by a simple-shear component parallel to strike of the thrust and a pure-shear component normal to it (Fig. 17). The instantaneous principal axes of shortening and extension parallel to the basal plane may be calculated (Walcott, 1978a; Fossen and Tikoff, 1993) and are plotted in Fig. 18(b). (The kinematic vorticity number (W_k) calculated from Fossen and Tikoff (1993, equation 21) for a net displacement direction similar to that on the basal Waikukupa thrust is less than 0.81, so that the axis of maximum elongation is in fact normal to the basal plane—the elongation axis plotted is the

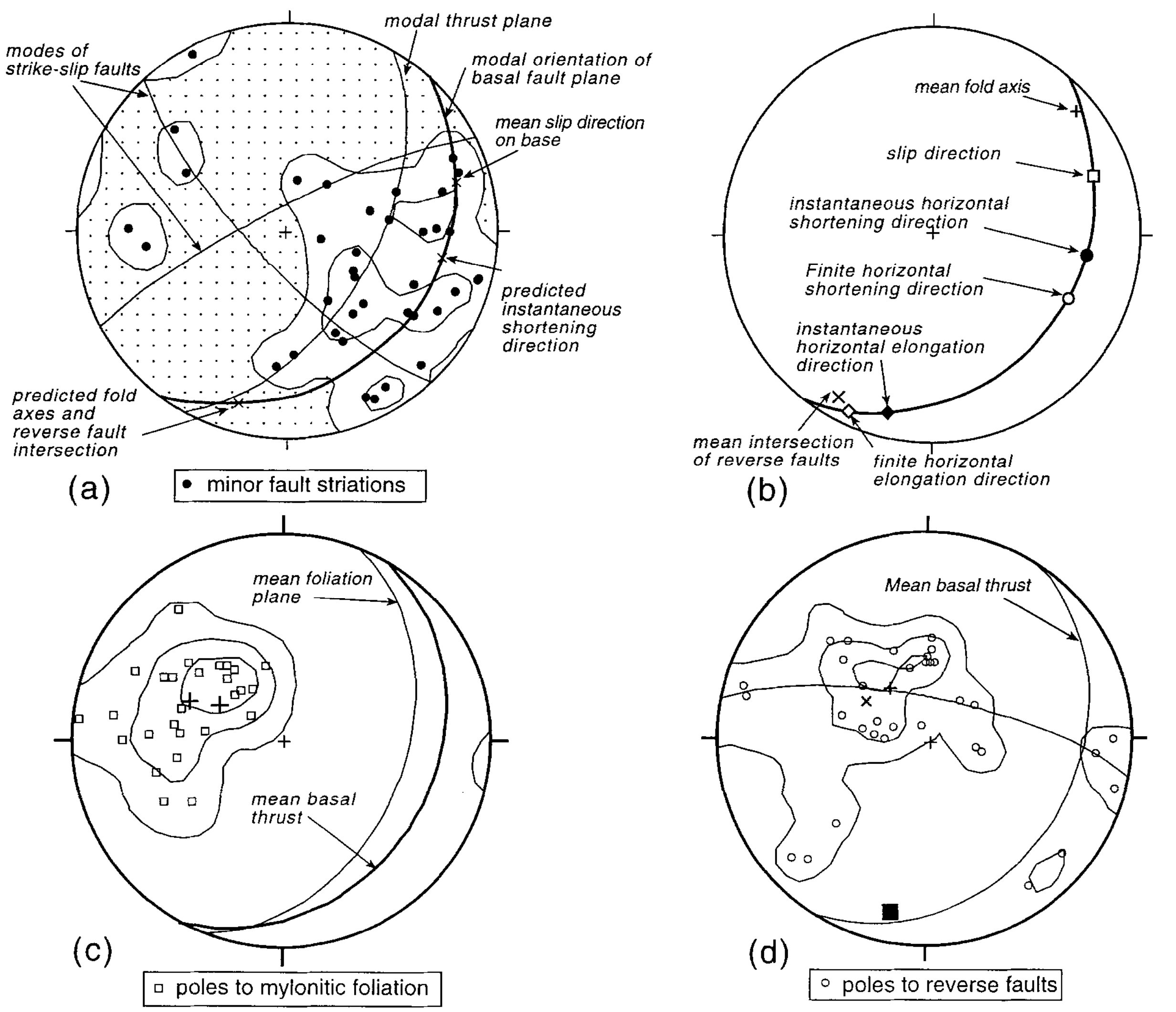

Fig. 18. (a) All striations (filled circles, 37 points) on minor reverse and strike-slip faults, Waikukupa thrust. Major modes of fault-plane orientations (one thrust and two strike-slip) are shown as great circles. The mean orientation of the basal fault plane is shown as a bold great circle. (b) Calculated axes of instantaneous and finite shortening and elongation parallel to the basal Waikukupa thrust, using the method of Fossen and Tikoff (1993) for the transpressional deformation of a block bounded by the basal thrust and subject to net convergence parallel to the basal striations. Also plotted are the mean intersection of reverse faults (from Fig. 12b) and the mean fold axis (from Fig. 12a). See text for discussion. (c) Poles to mylonitic foliation (open squares, 27 points) in the Hare Mare thrust sheet. Mean poles to foliation and to basal thrust are shown as crosses. (d) Poles to reverse faults (open circles, 31 points; upright cross, mean pole), Hare Mare thrust sheet. The best-fit girdle has its axis plunging gently south (filled square). The mean basal thrust plane (St Andrews cross, pole) is also plotted. Contours are in intervals of 3 x uniform.

intermediate strain axis.) During deformation, the principal axes rotate with increasing strain (Fossen and Tikoff, 1993). The apparent shortening of foliation estimated from down-dip sections through the Waikukupa thrust (see above) is 15–20%; the predicted orientations of the rotated axes of finite strain, assuming this shortening value approximates the principal contraction, are also plotted in Fig. 18(b) (calculated according to Fossen and Tikoff, 1993).

The orientations of structures within the Waikukupa thrust sheet are broadly consistent with these calculations. The mean intersection of reverse fault planes, which would be expected to correspond to the principal elongation direction within the basal plane, plots fairly close to the predicted position (Fig. 18b). Figure 18(a) shows striations from both reverse and strike-slip faults together with modal fault planes. Most striations plot within the SE quadrant and suggest a shortening direction of ESE, consistent with the predicted axis of shortening. The mean foliation fold axis and measured hinges are approximately parallel to the strike of the thrust (Fig. 12a), which is not quite as predicted by the model. On the other hand, poles to foliation and reverse faults within the mylonites of the Hare Mare thrust fall on a girdle with an axis plunging gently to the south, compatible with the simple-strain models of transpression (Fig. 18c & d).

The axis of folding within the Waikukupa thrust sheet appears to be more controlled by the strike of the foliation than by the predicted strain axes. This difference may not be significant given the noise in the data (Fig. 12a). On the other hand, it may reflect the failure of a simple model based on homogeneous strain to account for the mechanics of deformation of a wedge with an upper surface (i.e. the ground surface) which is neither planar nor parallel to the base.

The shape of the upper surface has two principal effects. In a critical wedge, the principal axes of stress and instantaneous strain are not parallel and perpendicular to the basal thrust but are rotated towards parallel and normal to the free surface (Davis *et al.*, 1983; Dahlen, 1984; Koons, 1994). The axis of folding will be closer to the line of intersection of the foliation with the free surface than in the simple block model.

Secondly, the effect of a non-planar upper surface introduces mechanical inhomogeneities and makes any treatment of internal deformation in terms of homogeneous transpression questionable. Modelling by Robin and Cruden (1994), assuming no slip on the lateral boundaries of the block, results in an inhomogeneous strain with strain axes curving into parallelism with the block margins. In sandbox experiments of wedge evolution by Mulugeta and Koyi (1992), episodic development occurs as accretion at the toe renders the wedge subcritical and is followed by internal shortening. Shortening initially occurs, at less than 20% strain, by development of kink folds which localize into shears at higher strains. The point of nucleation of the kinks is at a 'sticking' point in the wedge. In the case of the Waikukupa thrust, subcriticality is due to erosion rather than accretion but, in a similar way to the accretion experiments, it is not evenly distributed over the whole wedge. Hence, the folding may have occurred sequentially in space and be controlled by the position and orientation of the valley (block 'boundary') as it cut through the wedge. Certainly, critical wedge theory would predict that irregularities in the shape of the upper surface will affect the location of internal deformation (e.g. Stockmal, 1983).

The internal deformation of both the Hare Mare and Waikukupa thrust sheets is, nevertheless, broadly compatible geometrically with simple block models of transpressional deformation despite the undoubted mechanical complications introduced by surface geometry.

IMPLICATIONS FOR TRANSPRESSIONAL FAULT ZONE EVOLUTION

Although the fault geometry described above has some aspects in common with thrust complexes (e.g. Boyer and Elliott, 1982), and its evolution may be compared with that of a classic fold–thrust belt, it differs in several respects. It is of relatively small size, forms part of an oblique-slip fault zone with high rates of slip, and is subject to both rapid erosion and aggradation during late Quaternary glaciations.

Localized thrusting along the Alpine Fault was described briefly by Wellman (1955) and explained by gravity collapse of the range front along a vertical strike-slip fault. Sylvester and Smith (1976) described folding and thrusting on the southern San Andreas Fault, which they concluded results from shortening of the crust (transpression) between two closely spaced, vertical strike-slip faults and its displacement outwards over the margins of the squeezed block. We argue that the structures we describe here and similar structures elsewhere along the Alpine Fault (Cooper and Norris, 1994; Simpson *et al.*, 1994; Norris and Cooper, 1995) are an integral part of a major transpressional fault zone and result directly from the oblique shear within the zone.

The internal structure of the thrust sheets is indicative of shortening, except at the leading edge where the basal thrust bends onto the horizontal ground surface. The basal thrust surfaces dip at 20–30° and cut up through the originally overlying gravel sequences. The deep-seated cataclasite and mylonite sequence is emplaced along the thrusts indicating that the latter are extensions of the main Alpine Fault plane and root back onto it at depth. Slip directions on the basal thrusts are subparallel to the overall oblique fault movement. All these features argue against gravity collapse of the range front as the mechanism of emplacement.

The central section of the Alpine Fault consists of oblique thrust segments offset and linked by strike-slip segments (Fig. 1) (Norris and Cooper, 1995). The slip direction on all segments is similar, and subparallel to the plate vector. Coupled with the detailed observations summarized above, this pattern is strong evidence that the oblique thrust segments are an integral part of the plate-boundary deformation and are driven by interplate movement. This style of partitioning of displacement along the boundary onto thrust systems linked by en échelon strike-slip faults we here term 'serial partitioning' (Fig. 19a). Other possible examples of this structural style are described near San Francisco by Aydin and Page (1984) and in southern California by Bilham and Williams (1985).

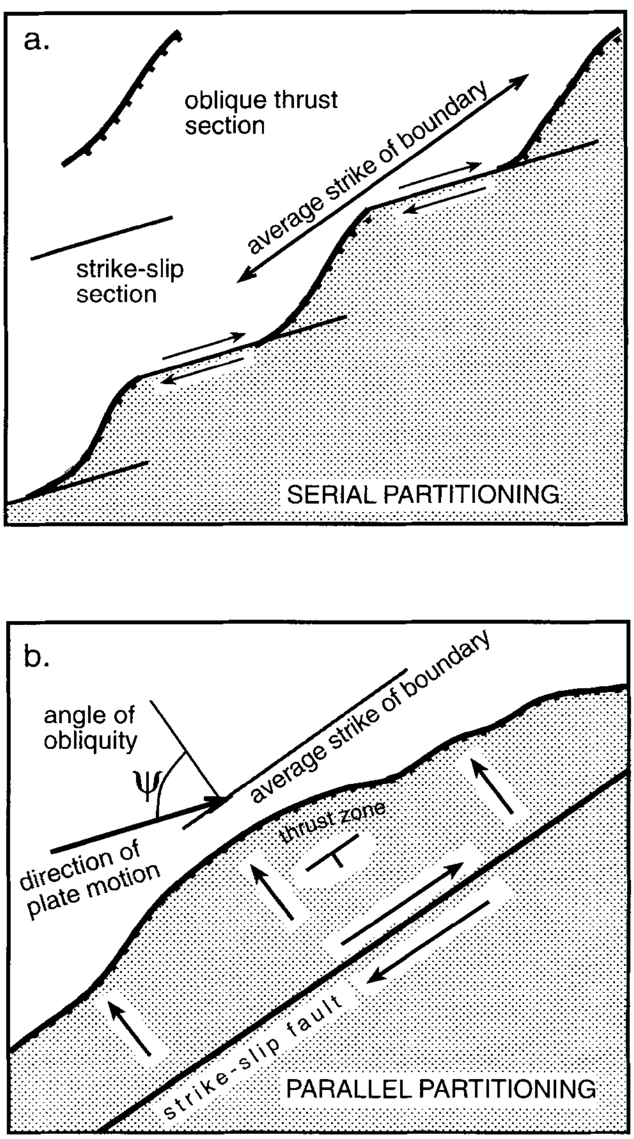

Fig. 19. Schematic diagrams showing two types of displacement partitioning on a transpressional fault: (a) *serial partitioning*, with the formation of linked strike-slip and thrust segments; (b) *parallel partitioning*, with the formation of parallel zones of strike-slip and thrusting. This is similar to the partitioning model investigated by McCaffrey (1992). The angle of obliquity of convergence, ψ (used in eqns 2 and 3), is illustrated. See text for discussion.

Mechanical models of the Southern Alps require rapid erosion on the western side to localize the major zone of displacement (Koons, 1990; Beaumont *et al.*, 1992). The straightness of the Alpine Fault overall is striking and perhaps unusual for a convergent structure. Rapid erosion and the efficient removal of detritus by the Tasman Sea, however, prevents the development of extensive low-angle thrusts to the west, maintaining the straight trace. The Waikukupa section clearly illustrates this process. Thrust complexes extending westward lose their mechanical integrity by erosion, causing 'out of sequence' imbrication. Because erosion is not uniform but controlled by major rivers (e.g. Koons, 1989), rearward imbrication is also localized, leading directly to a serially partitioned range front (Fig. 15). Thus, while the en échelon pattern of thrust and strike-slip portions of the fault may be characteristic of transpression (e.g. Aydin and Nur, 1985; Koons, 1994), in detail the pattern is strongly influenced by erosion (Norris and Cooper, 1995).

The minimum rate of downcutting ($D_{\min}$) required to cause rearward imbrication must be greater than the rate of vertical uplift; i.e.

$$D_{\min} > d_s \sin \beta \qquad (1)$$

where d_s is the dip-slip component on the fault and β is the dip of fault. For the Waikukupa thrust, the uplift rate is about 5 mm/year, while the estimated rate of downcutting by the Waikukupa River has averaged around 10 mm/year since the Okarito glacial advance.

Active growth of thrust sheets will tend to occur when and where they are protected from erosion, for instance immediately following a glacial period when thick deposits of moraine are piled up along the range front. Conversely, rearward imbrication and abandonment will occur at times of rapid erosional downcutting, restoring the fault trace to its original location (Wellman, 1955, makes a suggestion along these lines). Thus, there is likely to be a climatic influence on the development of range front structures and, in fact, on the shape and character of the range front. This is portrayed here by the growth and abandonment of the Waikukupa thrust in the period between the Okarito and Moana glaciations. In systems with lower overall erosion rates than the Southern Alps, such coupling may be even more pronounced.

An alternative and apparently common form of displacement partitioning is where the strike-slip component is accommodated on a separate, steeply dipping fault zone behind and striking parallel to the outward moving thrust sheets so that the displacement is partitioned onto parallel structures. We here term this style of displacement partitioning 'parallel partitioning' (Fig. 19b). Many examples of this have been described in the literature, particularly from the San Andreas Fault (e.g. Sylvester and Smith, 1976; Mount and Suppe, 1987; Namson and Davis, 1990). Several authors (e.g. Tikoff and Teyssier, 1994; Braun and Beaumont, 1995) have argued that the major factor in determining whether parallel partitioning will occur is the angle of obliquity of convergence. Undoubtedly that is of importance, but here we suggest another factor that may be significant in developing parallel partitioning along a transpressional fault zone.

In the New Zealand case, not only is erosion rapid on the west side of the Southern Alps, but the Tasman Sea a few kilometres farther west forms a large sink for all the detritus, which only builds up during glaciations. According to the models of Koons (1990) and Beaumont *et al.* (1992), this efficient erosion, transport and disposal system maintains the Alpine Fault as the principal locus of deformation. As we argue above, in areas where erosion is less and/or efficient transport is lacking, so that detritus builds up adjacent to the fault, oblique thrusts such as observed here will become more extensive. Failure of the footwall is likely to occur (e.g. Panian and Pilant, 1990) producing a more conventional forward-propagating system. Early stages of this are seen in the deformation and thrusting of the fluvioglacial gravels in the present study, and at deeper levels in the development of thrusts within basement west of the fault (Rattenbury, 1986). In order for the thrust wedge to grow, thickening by faulting and folding within the wedge is required (Morley, 1988; Mulugeta and Koyi, 1992). Thus, a system of active folds and thrusts within a wedge of increasing height, depth and length will evolve.

As these thrust structures are oblique to the slip vector, however, it is questionable as to how far they can extend away from the main fault zone as low-dipping structures and still retain mechanical integrity and fully transpressional features. McCaffrey (1992) has formulated the force balance between slip on a basal thrust and slip on a vertical strike-slip fault at the back of an accretionary wedge (e.g. Fig. 19b) and gives the following relationship (after McCaffrey, 1992, equation 5):

$$\sin(\psi_{\max}) = \sin(\delta)\tau_s'/\tau_t' \qquad (2)$$

where $\psi_{\max}$ is the critical angle of obliquity of convergence at which slip occurs on the strike-slip fault, δ is the angle of dip of the thrust fault, and τ'_s and τ'_t are the critical average resolved shear stresses for slip on the strike-slip fault and thrust, respectively. McCaffrey's equation ignores topography; for a thrust wedge with a surface slope of angle α and basal slope β, the equation becomes:

$$\sin(\psi_{\max}) = \cos\beta(\tan\alpha + \tan\beta)\tau_s'/\tau_t'. \qquad (3)$$

τ'_s is itself partly dependent on the angle of convergence through the resolved normal stress on the strike-slip fault. Even if this is factored into the calculations, however, $\psi_{\max}$ is still a function of the average dip of the basal thrust. As this becomes smaller, the degree of obliquity at which partitioning occurs also becomes less. As the fold–thrust complex along the transpressional range front develops and extends outwards, the dip of the base becomes less, making partitioning of the parallel component onto a vertical strike-slip fault more likely. If erosion is effective in preventing the development of a low-angle thrust complex, parallel partitioning will be less likely to occur for a given angle of obliquity and instead the range front will become partitioned in a serial fashion (Fig. 19a).

Critical wedge models (Davis *et al.*, 1983; Dahlen, 1984; Koons, 1994) are useful in understanding the behaviour of the fault zone deformation, in particular the growth, deformation and abandonment of the thrust sheets. The detailed behaviour of the small-scale structures described here at the inboard toe of the Southern Alps is governed by the same mechanical principles as the deformation of the whole mountain range. Feedback between tectonic processes of deformation and surface processes of erosion may be significant at the scale of the mountain range (Koons, 1990; Beaumont *et al.*, 1992) and at the scale of a fault zone and river valley.

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REFERENCES

Adams, C. J. (1981) Uplift rates and thermal structure in the Alpine Fault Zone and Alpine Schists, Southern Alps, New Zealand. In *Thrust and Nappe Tectonics*, eds K. McClay and N. J. Price, pp. 211–222. Geological Society of London Special Publication **9**.

Allis, R. G. (1986) Mode of crustal shortening adjacent to the Alpine Fault, New Zealand. *Tectonics* **5**, 15–32.

Aydin, A. and Nur, A. (1985) The types and role of stepovers in strike-slip tectonics. In *Strike-slip Deformation, Basin Formation, and Sedimentation*, eds K. T. Biddle and N. Christie-Blick, pp. 35–44. Society of Economic Paleontologists and Mineralogists Special Publication **37**.

Aydin, A. and Page, B. M. (1984) Diverse Pliocene–Quaternary tectonics in a transform environment. *Bulletin of the Geological Society of America* **95**, 1303–1317.

Beaumont, C., Fullsack, P. and Hamilton, J. (1992) Erosional control of active compressional orogens. In *Thrust Tectonics*, ed K. R. McClay, pp. 1–18. Chapman and Hall, London.

Berryman, K. R., Beanland, S., Cooper, A. F., Cutten, H. N., Norris, R. J. and Wood, P. R. (1992) The Alpine Fault, New Zealand: variation in Quaternary tectonic style and geomorphic expression. *Annales Tectonicae* **VI**, 126–163.

Bibby, H. M., Haines, A. J. and Walcott, R. I. (1986) Geodetic strain and the present day plate boundary zone through New Zealand. In *Recent Crustal Movements of the Pacific Region*, eds W. I. Reilly and B. E. Harford, pp. 427–438. Royal Society of New Zealand Bulletin **24**.

Bilham, R. and Williams, P. (1985) Sawtooth segmentation and deformation processes on the southern San Andreas fault, California. *Geophysical Research Letters* **12**, 557–560.

Boyer, S. E. and Elliott, D. (1982) Thrust Systems. *Bulletin of the American Association of Petroleum Geologists* **66**, 1196–1230.

Braun, J. and Beaumont, C. (1995) Three-dimensional numerical experiments of strain partitioning at oblique plate boundaries: Implications for contrasting tectonic styles in the southern Coast Ranges, California, and central South Island, New Zealand. *Journal of Geophysical Research* **100**, 18059–18074.

Brown, E. H. and Talbot, J. L. (1989) Orogen-parallel extension in the North Cascades crystalline core. *Tectonics* **8**, 1105–1114.

Bull, W. B. and Cooper, A. F. (1986) Uplifted marine terraces along the Alpine fault, New Zealand. *Science* **234**, 1225–1228.

Chapple, W. M. (1978) Mechanics of thin-skinned fold-and-thrust belts. *Bulletin of the Geological Society of America* **89**, 1189–1198.

Cobbold, P. R., Gapais, D. and Rosello, E. A. (1991) Partitioning of transpressive motions within a sigmoidal foldbelt: the Variscan Sierras Australes, Argentina. *Journal of Structural Geology* **13**, 777–786.

Cooper, A. F. (1980) Retrograde alteration of chromium kyanite in metachert and amphibolite whiteschist from the Southern Alps, New Zealand, with implications for uplift on the Alpine Fault. *Contributions to Mineralogy and Petrology* **75**, 153–164.

Cooper, A. F. and Norris, R. J. (1994) Anatomy, structural evolution and slip rate of a plate-boundary thrust: The Alpine fault at Gaunt Creek, Westland, New Zealand. *Bulletin of the Geological Society of America* **106**, 627–633.

Cooper, A. F. and Norris, R. J. (1995) Displacement on the Alpine Fault at Haast River, South Westland, New Zealand. *New Zealand Journal of Geology and Geophysics* **38**, 509–514.

Dahlen, F. A. (1984) Non-cohesive critical Coulomb wedges: an exact solution. *Journal of Geophysical Research* **89**, 10125–10133.

Davis, D., Suppe, J. and Dahlen, F. A. (1983) Mechanics of fold-and-thrust belts and accretionary wedges. *Journal of Geophysical Research* **88**, 1153–1172.

DeMets, C., Gordon, R. G., Argus, D. F. and Stein, S. (1990) Current plate motions. *Geophysical Journal International* **101**, 425–478.

Elliott, D. (1976) The energy balance and deformation mechanisms of thrust sheets. *Philosophical Transactions of the Royal Society of London* **A283**, 289–312.

Fitch, T. J. (1972) Plate convergence, transcurrent faults, and internal deformation adjacent to southeast Asia and the western Pacific. *Journal of Geophysical Research* **77**, 4432–4460.

Fossen, H. and Tikoff, B. (1993) The deformation matrix for simultaneous simple shearing, pure shearing and volume change, and its application to transpression–transtension tectonics. *Journal of Structural Geology* **15**, 413–422.

Grapes, R. H. and Watanabe, T. (1994) Mineral composition variation in Alpine Schist, Southern Alps, New Zealand: Implications for recrystallization and exhumation. *The Island Arc* **3**, 163–181.

Grapes, R. I. (1993) Alpine fault gouge or "the green rock". *Geological Society of New Zealand Newsletter* **100**, 46–49.

Grapes, R. and Watanabe, T. (1992) Metamorphism and uplift of the Alpine Schist in the Franz Josef–Fox Glacier area of the Southern Alps, New Zealand. *Journal of Metamorphic Geology* **10**, 171–180.

Griffiths, G. A. and McSaveney, M. J. (1983) Distribution of mean annual precipitation across some steepland regions of New Zealand. *New Zealand Journal of Science* **26**, 197–209.

Hansen, V. L. (1989) Structural and kinematic evolution of the Teslin suture zone, Yukon: record of an ancient transpressional margin. *Journal of Structural Geology* **11**, 717–733.

Hanson, C. R., Norris, R. J. and Cooper, A. F. (1990) Regional fracture patterns east of the Alpine Fault between the Fox and Franz Josef glaciers, Westland, New Zealand. *New Zealand Journal of Geology and Geophysics* **33**, 617–622.

Harland, W. (1971) Tectonic transpression in Caledonian Spitzbergen. *Geological Magazine* **108**, 27–42.

Holdsworth, R. E. and Strachan, R. A. (1991) Interlinked system of ductile strike-slip and thrusting formed by Caledonian sinistral transpression in northeastern Greenland. *Geology* **19**, 510–513.

Holm, D. K., Norris, R. J. and Craw, D. (1989) Brittle/ductile deformation in a zone of rapid uplift: central Southern Alps, New Zealand. *Tectonics* **8**, 153–168.

Jamison, W. R. (1992) Stress controls on fold thrust style. In *Thrust Tectonics*, ed. K. R. McClay, pp. 155–164. Chapman and Hall, London.

Johnstone, R. D., Craw, D. and Rattenbury, M. S. (1990) Southern Alps Cu–Au hydrothermal system, Westland, New Zealand. *Mineralium Deposita* **25**, 118–125.

Koons, P. O. (1987) Some thermal and mechanical consequences of rapid uplift: an example from the Southern Alps, New Zealand. *Earth and Planetary Science Letters* **86**, 307–319.

Koons, P. O. (1989) The topographic evolution of collisional mountain belts: a numerical look at the Southern Alps, New Zealand. *American Journal of Science* **289**, 1041–1069.

Koons, P. O. (1990) Two-sided orogen: Collision and erosion from the sandbox to the Southern Alps, New Zealand. *Geology* **18**, 679–683.

Koons, P. O. (1994) Three-dimensional critical wedges: Tectonics and topography in oblique collisional orogens. *Journal of Geophysical Research* **99**, 12301–12315.

Landis, C. A. and Coombs, D. S. (1967) Metamorphic belts and orogenesis in southern New Zealand. *Tectonophysics* **4**, 501–518.

McCaffrey, R. (1991) Slip vectors and stretching of the Sumatran forearc. *Geology* **19**, 881–884.

McCaffrey, R. (1992) Oblique plate convergence, slip vectors, and forearc deformation. *Journal of Geophysical Research* **97**, 8905–8915.

Meisling, K. E. and Weldon, R. J. (1989) Late Cenozoic tectonics of the northwestern San Bernardino Mountains, southern California. *Bulletin of the Geological Society of America* **101**, 106–128.

Molnar, P. (1992) Brace–Goetze strength profiles, the partitioning of strike-slip and thrust faulting at zones of oblique convergence, and the stress–heat flow paradox of the San Andreas Fault. In *Fault Mechanics and Transport Properties of Rocks*, eds B. Evans and T.-F. Wong, pp. 435–460. Academic Press, London.

Morley, C. K. (1988) Out-of sequence thrusts. *Tectonics* **7**, 539–561.

Morrow, C. A., Shi, L. Q. and Byerlee, J. D. (1982) Strain hardening and strength of clay-rich fault gouges. *Journal of Geophysical Research* **87**, 6771–6780.

Mount, V. S. and Suppe, J. (1987) State of stress near the San Andreas fault: implications for wrench tectonics. *Geology* **15**, 1143–1146.

Mulugeta, G. and Koyi, H. (1992) Episodic accretion and strain partitioning on a model sand wedge. *Tectonophysics* **202**, 319–333.

Namson, J. S. and Davis, T. L. (1988) Seismically active fold and thrust belt in the San Joachim Valley, central California. *Bulletin of the Geological Society of America* **100**, 257–273.

Namson, J. S. and Davis, T. L. (1990) Late Cenozoic fold and thrust belt of the Southern Coast Ranges and Santa Maria Basin, California. *Bulletin of the American Association of Petroleum Geologists* **74**, 467–492.

Norris, R. J. and Cooper, A. F. (1995) Origin of small-scale segmentation and transpressional thrusting along the Alpine fault, New Zealand. *Bulletin of the Geological Society of America* **107**, 231–240.

Norris, R. J., Koons, P. O. and Cooper, A. F. (1990) The obliquely-convergent plate boundary in the South Island of New Zealand: implications for ancient collision zones. *Journal of Structural Geology* **12**, 715–725.

Oldow, J., Bally, A. and Avé Lallemant, H. (1990) Transpression, orogenic float, and lithospheric balance. *Geology* **8**, 991–994.

Panian, J. and Pilant, W. (1990) A possible theoretical explanation for foreland thrust propogation. *Journal of Geophysical Research* **95**, 8609–8615.

Pearson, C. (1994) Geodetic strain determinations from the Okorito and Godley–Tekapo regions, central South Island, New Zealand. *New Zealand Journal of Geology and Geophysics* **37**, 309–318.

Platt, J. P. (1993) Mechanics of oblique convergence. *Journal of Geophysical Research* **98**, 16239–16256.

Ratschbacher, L. (1986) Kinematics of Austro-Alpine cover nappes: changing translation paths due to transpression. *Tectonophysics* **125**, 335–356.

Rattenbury, M. S. (1986) Late low angle thrusting and the Alpine Fault, central Westland, New Zealand. *New Zealand Journal of Geology and Geophysics* **29**, 437–446.

Reed, J. J. (1964) Mylonites, cataclasites and associated rocks along the Alpine Fault, South Island, New Zealand. *New Zealand Journal of Geology and Geophysics* **7**, 654–684.

Reyners, M. (1987) Subcrustal earthquakes in the central South Island, New Zealand, and the root of the Southern Alps. *Geology* **15**, 1168–1171.

Robin, P.-Y. F. and Cruden, A. R. (1994) Strain and vorticity patterns in ideally ductile transpression zones. *Journal of Structural Geology* **16**, 447–466.

Sanderson, D. J. and Marchini, W. R. D. (1984) Transpression. *Journal of Structural Geology* **6**, 449–458.

Sibson, R. H., White, S. H. and Atkinson, B. K. (1979) Fault rock distribution and structure within the Alpine Fault Zone: a preliminary account. In *The Origin of the Southern Alps*, eds R. I. Walcott and M. M. Cresswell, pp. 55–65. Bulletin of the Royal Society of New Zealand **18**.

Simpson, G. D. H., Cooper, A. F. and Norris, R. J. (1994) Late Quaternary evolution of the Alpine fault zone at Paringa, South Westland. *New Zealand Journal of Geology and Geophysics* **37**, 49–58.

Stockmal, G. S. (1983) Modelling of large-scale accretionary wedge deformation. *Journal of Geophysical Research* **88**, 8271–8287.

Suggate, R. P. (1963) The Alpine fault. *Transactions of the Royal Society of New Zealand* **2**, 105–129.

Suggate, R. P. (1990) Late Pliocene and Quaternary glaciations of New Zealand. *Quaternary Science Reviews* **9**, 175–197.

Sutherland, R. (1994) Displacement since the Pliocene along the southern section of the Alpine fault, New Zealand. *Geology* **22**, 327–331.

Sylvester, A. G. and Smith, R. R. (1976) Tectonic transpression and basement-controlled deformation in the San Andreas fault zone, Salton trough, California. *Bulletin of the American Association of Petroleum Geologists* **60**, 2081–2102.

Tikoff, B. and Teyssier, C. (1994) Strain modeling of displacement-field partitioning in transpressional orogens. *Journal of Structural Geology* **16**, 1575–1588.

Tippett, J. M. and Kamp, P. J. J. (1993) Fission track analysis of the late Cenozoic vertical kinematics of continental Pacific crust, South Island, New Zealand. *Journal of Geophysical Research* **98**, 16119–16148.

Walcott, R. I. (1978a) Present tectonics and late Cenozoic evolution of New Zealand. *Geophysical Journal of the Royal Astronomical Society* **52**, 137–164.

Walcott, R. I. (1978b) Geodetic strains and large earthquakes in the axial tectonic belt of North Island, New Zealand. *Journal of Geophysical Research* **83**, 19–29.

Walcott, R. I. (1996) Contemporary relative motions of the Pacific and Australian plates. *Geological Society of New Zealand Newsletter* **110**, 48.

Warren, G. (1967) Sheet 17 Hokitika. In *Geological Map of New Zealand* 1:250 000 (1st edn). Department of Scientific and Industrial Research, Wellington, New Zealand.

Wellman, H. W. (1955) *The Geology Between Bruce Bay and Haast River, South Westland.* New Zealand Geological Survey Bulletin n.s. **48** (2nd edn).

Willemin, J. H. (1984) Erosion and the mechanics of shallow foreland thrusting. *Journal of Structural Geology* **6**, 425–432.

Wood, P. R. and Blick, G. H. (1986) Some results of geodetic fault monitoring in South Island, New Zealand. In *Recent Crustal Movements of the Pacific Region*, eds W. I. Reilly and B. E. Harford, pp. 39–45. Royal Society of New Zealand Bulletin **24**.

Woodward, D. J. (1979) The crustal structure of the Southern Alps, New Zealand, as determined by gravity. In *The Origin of the Southern Alps*, eds R. I. Walcott and M. M. Cresswell, pp. 95–98. Royal Society of New Zealand Bulletin **18**.

Zoback, M. D., Zoback, M. L., Mount, V. S., Suppe, J., Eaton, J. P., Healy, J. H., Oppenheimer, D., Reasenberg, P., Jones, L., Raleigh, C. B., Wong, I. B., Scotti, O. and Wentworth, C. (1987) New evidence on the state of stress of the San Andreas fault system. *Science* **238**, 1105–1111.