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The obliquely-convergent plate boundary in the South Island of New Zealand: implications for ancient collision zones

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Abstract—The Alpine Fault of New Zealand forms the western boundary of a zone of distributed deformation formed by the oblique convergence of continental crust belonging to the Pacific and Australian plates. Structural and geodetic data from the Alpine Fault show that a large proportion of the total plate displacement is accommodated by rapid oblique slip on the fault. The remaining displacement is distributed over a 200 km wide zone to the east. The collision may be modelled as a two-sided deforming orogen, with the partitioning of deformation being controlled by erosional differences between the narrow high-strain inboard and broad low-strain outboard sides. In ancient collision zones, little evidence may remain of the nature and amount of displacement on the inboard side. Partitioning of deformation among pre-existing structures complicates interpretation of the outboard zone. Radiometric ages may post-date collision by several tens of millions of years and indicate slow isostatic uplift and unroofing. Collision ages, if preserved, may be recognized by high uplift rates calculated from muscovite–biotite pairs.

INTRODUCTION

The concept of tectonostratigraphic terranes, introduced mainly through work in western North America (Coney *et al.* 1980), brought two major reorientations to the geological view of orogenic belts. The recognition of far-travelled allochthonous terranes necessitates caution in interpreting palaeographies of orogenic belts in terms of the present distribution of tectonically-bounded segments—the original spatial relationships of such segments must be left open until clear evidence is obtained. This requires the objective subdivision of orogenic belts, which is at the heart of the terrane concept. The second major change in outlook is the greater significance attached to three dimensions in orogenic models. Traditional models tended to assume that orogenic belts are cylindrical structures in which the important displacements are normal to the trend of the belt and therefore may be summarized in a simple cross-section (e.g. Dewey & Bird 1970, Coombs *et al.* 1976). The terrane model emphasizes significant components of strike-slip parallel to the belt which may be more significant than the orthogonal convergence.

Since the introduction of the terrane concept, many studies have attempted to subdivide various orogens around the world into tectonostratigraphic terranes (e.g. Howell 1985). In addition, detailed structural investigations of terrane boundaries have been made with the aim of shedding light on the displacement history (e.g. Basden *et al.* 1987, Offler & Williams 1987). Geological and geochronological data from terranes have been used to determine the timing of accretion, etc. (e.g. reviews by Howell *et al.* 1985, Packham 1987, Zartman 1988). In order to interpret data reliably, an understanding is needed of the tectonics of modern collision zones, particularly those in which the collision is oblique. An example of such a zone is found in the South Island of New Zealand.

The Australian–Pacific plate boundary crosses the continental crust of the South Island as a zone of distributed deformation up to 200 km wide (Walcott 1978a). The principal locus of displacement between the two plates is the Alpine Fault (Fig. 1), a major strike-slip fault first recognized by Wellman (1953) as having 480 km of right-lateral slip since the Jurassic. With the advent of plate tectonics, it has generally been viewed as a transform fault linking the W-directed subduction zone in the North Island with an E-directed oblique subduction zone in the southwest of the South Island (e.g. Le Pichon 1968). More recent work has established the inception of the plate boundary in late Eocene–early Oligocene time and the development of the Alpine Fault as a through-going transform at around 25 Ma (Molnar *et al.* 1975, Carter & Norris 1976, Kamp 1986, Cooper *et al.* 1987). Displacement has become increasingly oblique over the last 10 Ma, resulting in around 70 km of shortening normal to the fault (Molnar *et al.* 1975, Walcott 1978a, Allis 1986). The situation is essentially one of oblique continental collision, with the present displacement vector trending at about 30° to the fault in the central South Island (Walcott 1979).

As part of an ongoing co-operative study of collision-zone tectonics in the Southern Alps (Koons *et al.* in preparation), we review some aspects of the structure of the plate boundary and the strain adjacent to it, and develop a model for the internal kinematics of the

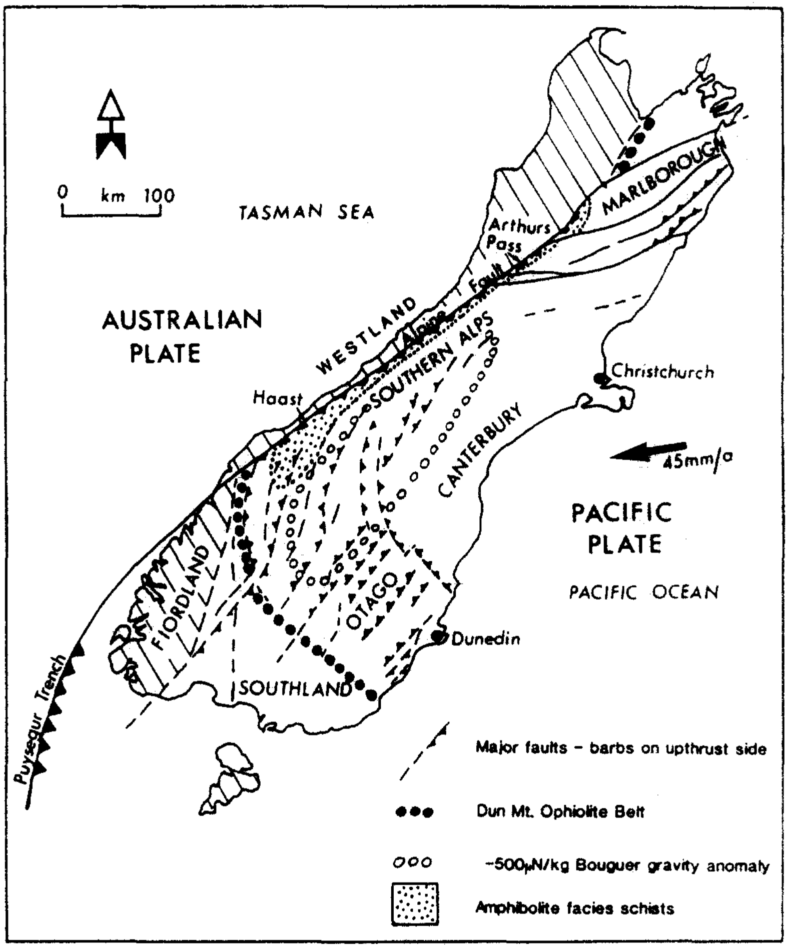

Fig. 1. Tectonic map of the South Island. The heavy arrow represents the direction of motion of the Pacific plate relative to the Australian plate, determined from the pole of rotation given by Chase (1978) (see also Walcott 1979).

oblique collision zone. Our conclusions have significance for the interpretation of ancient examples of similar zones of oblique collision.

THE SOUTH ISLAND COLLISION ZONE

The Alpine Fault is a major terrane boundary. To the west lies a terrane of early Palaeozoic low-grade metasediments intruded by late Palaeozoic granites and locally metamorphosed to amphibolite facies with the further intrusion of granite in the Cretaceous (Bishop *et al.* 1985, Rettenbury 1987), while to the east lies a terrane comprising late Palaeozoic and Mesozoic schists and low-grade metasediments. The current displacement vector of the Pacific and Australian plates in the central South Island may be determined from the pole of rotation of the two plates as established from oceanic magnetic anomalies (Molnar *et al.* 1975, Chase 1978, Walcott 1979, C. DeMets personal communication to C. Pearson 1988). Based on the Chase (1978) pole, the displacement vector has a trend of 084° and a rate of 45 mm a^{-1}. The Alpine Fault strikes 055°, so that the displacement may be resolved into components of 40 mm a^{-1} parallel to the fault and 22 mm a^{-1} normal to it (Walcott 1979). The Chase pole for the Australian–Pacific rotation gives the maximum rate and convergence angle of all the various poles reported in the literature (cf. Minster *et al.* 1974, Minster & Jordan 1978, C. DeMets personal communication to C. Pearson 1988); a recent redetermination by DeMets *et al.* (C. DeMets personal communication to C. Pearson 1988) gives a vector trending 071° and a rate of 39 mm a^{-1}, with components of 37 mm a^{-1} parallel and 11 mm a^{-1} normal to the fault. Plate tectonic reconstructions show that the total displacement rate and the degree of convergence have both increased with time (Walcott 1978a, Smith 1981), so that the current rates are the highest since the inception of the plate boundary. Deformation is not restricted to the fault itself but is distributed east of the fault trace over a zone at least 100 km wide (Walcott 1978a) and almost double this in Otago (Fig. 1). A result of this strain distribution has been a bending of features east of the fault, well displayed by the narrow Dun Mountain Ophiolite Belt (Norris 1979).

The most obvious effect of the convergence is the rapid uplift of the Southern Alps, an asymmetric mountain range with short steep gradients on the west and longer, more gradual, gradients to the east. The Main Divide reaches over 3000 m in the central South Island and runs approximately parallel to, and 10–20 km east of, the Alpine Fault. Data from immediately east of the fault indicate uplift rates approaching 10 mm a^{-1} (Wellman 1979, Bull & Cooper 1986).

The major metamorphism in the Haast Schists east of the Alpine Fault took place during the Mesozoic and the schists were uplifted and eroded to a low level surface by late Cretaceous. At this time, the area of highest metamorphic grade exposed at the surface, coinciding with the axis of maximum uplift, formed an elongate zone extending from the vicinity of Dunedin (Fig. 1), where garnet zone schists crop out beneath the Cretaceous unconformity (Craw 1981), northwest at least as far as Haast where the exposed rocks were probably of low greenschist facies. The NE-trending Alpine Fault crosses the earlier zone of uplift at a high angle. Late Cainozoic differential uplift on the fault has brought high-grade amphibolite-facies schists to the surface, forming a narrow strip immediately east of the fault (Fig. 1). The metamorphic grade is highest adjacent to the fault and decreases eastwards, with the peaks of the Main Divide being composed of low-grade Permo-Triassic metasediments.

Although now steeply dipping adjacent to the fault due to Cainozoic uplift, the metamorphic isograds and intervening assemblages are consistent with metamorphism under a 'normal' geothermal gradient, similar to schists further east in Otago (Cooper 1974, 1983, Grapes & Otsuki 1983). Synmetamorphic structures in the schists are also continuous with Mesozoic structures further east (Grindley 1963, Cooper 1974, Findlay 1987, Rattenbury *et al.* 1989), indicating that the main metamorphic assemblages and isograds were developed during regional metamorphism in the Mesozoic; they have subsequently been tilted and uplifted, with some overprinted deformation and retrogression, during the late Neogene convergence. The evidence is contradictory to the view, postulated by Scholz *et al.* (1979), that the high-grade Alpine Schists result from extensive frictional heating during movement on the Alpine Fault. In addition to the petrological evidence, the fact that the movement on the Alpine Fault is clearly Cainozoic whereas the main metamorphic zonation is Mesozoic precludes a genetic association between the two.

A large negative Bouguer gravity anomaly is associated with the Southern Alps (Fig. 1) (Reilly & Whiteford 1979). As pointed out by Woodward (1979) and Allis (1986), the axis of the anomaly does not underlie the Main Divide but is displaced to the east and trends in a more northerly direction. The presence of the anomaly implies a thickened crust due to development of a root. It is wider and larger at its southern end, and terminates abruptly north of the Dun Mountain Ophiolite Belt. The thickness of the crust forming the root as calculated from the gravity anomaly depends on the values taken for crustal density and the 'normal' thickness of the crust away from the root. Woodward (1979) calculated a maximum thickness of 50 km using a normal crust of 30 km whereas Allis (1986) managed to reduce this to 35 km by assuming a thinner (25 km) original crust.

Late Cainozoic reverse faults are widespread east of the Alpine Fault, their development synchronous with the increase in the component of convergence on the plate boundary during the late Miocene. Many represent reactivated Cretaceous normal faults (e.g. Mutch & Wilson 1952, Bishop 1974, Turnbull *et al.* 1975), originating during probable lithospheric stretching associated with the opening of the Tasman Sea and southwest Pacific (Spörli 1980). The faults strike generally either NE or NW, although the northwest faults tend to bend into a northerly direction on approaching the Alpine Fault (Fig. 1) (Norris 1979). The amount of dip-slip on these faults during the late Cainozoic ranges from a few hundred metres to several kilometres, although the strike-slip component, while probably significant, is generally unknown (Bishop 1974, Makgill & Norris 1983, Beanland *et al.* 1986). The distribution of NW- and NE-trending faults is uneven over the area, with regions being dominated by one set or the other (Fig. 1). The southern margin of the schists in northwest Otago is marked by the Dun Mountain Ophiolite Belt, bounded by and parallel to major NW-trending faults. South of here, the crystalline Fiordland block lies adjacent to the plate boundary and is bounded by major zones of Cainozoic deformation (Norris & Carter 1982). This deformation includes at least 20 km of right-lateral slip along the eastern margin (Norris & Carter 1980, Constantine 1987), with corresponding shortening, bending and basin eversion in northern Fiordland and adjacent areas (Norris & Carter 1982).

The Alpine Fault

On the 1:1,000,000 map (N.Z. Geological Survey 1972), the Alpine Fault runs essentially as a single straight structure from Arthur's Pass in the north to Fiordland in the south. Further north, it splits into several faults of the Marlborough system whereas off Fiordland, convergence is redistributed west of the fault where ocean floor of the Tasman Sea is being obliquely subducted (Davey & Smith 1983).

In detail, the Alpine Fault is not straight, but consists of more northerly-trending zones dominated by oblique overthrusting linked by faults of a more strike-slip character (Fig. 2). The conclusions in this section will be documented more fully by Norris & Cooper (work in preparation). The hanging walls of the overthrusts expose sections through a highly crushed basal cataclasite grading up through fractured mylonites into fairly intact mylonites and protomylonites derived at depth from the overlying schists (Fig. 2) (Sibson *et al.* 1979, Norris *et al.* 1987). The basal contact in the thrust segments generally dips at about 30°E; the overlying mylonites are highly imbricated near to the base but further east dip fairly consistently at around 50°E (Sibson *et al.* 1979, Norris *et al.* 1987). The mylonite zone varies in thickness, due in part to dismemberment during brittle faulting, but reaches thicknesses of up to 1 km. Slickenside lineations within the basal cataclasite generally trend around 080°, as do the strike-slip transfer faults (Fig. 2), indicating displacement parallel to the overall plate velocity vector. Stretching lineations within the mylonites also trend parallel to the plate vector (Sibson *et al.* 1979).

The thickness of the uplifted mylonites compared with

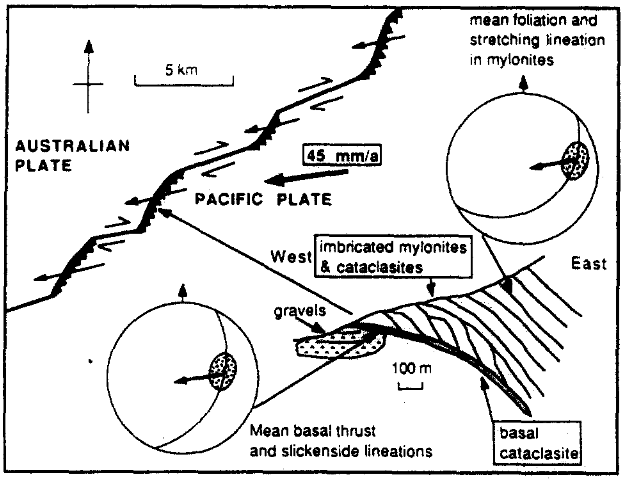

Fig. 2. Diagrammatic representation of a section of the Alpine Fault in central Westland, showing segmentation into thrust (barbs on upthrust side) and strike-slip portions. Cross-section is of a typical thrust segment; form lines represent mylonitic foliation. (Mylonite orientation data summarized from Sibson *et al.* 1979. Rest based on unpublished work by Norris & Cooper).

the cataclasites suggests that the Alpine Fault widens at depth as a ductile mylonite zone striking 055° and dipping around 50°E. Displacement within the zone is oblique and parallel to the plate vector. Nearer the surface, the deformation becomes brittle, the principal zone of displacement narrows, and the fault segments into thrust and strike-slip sections. The net-slip direction remains the same, however. The rate of displacement on the fault itself has proved difficult to determine due to the lack of preserved, well-dated, displaced features (cf. Wellman 1953). Some recent results based on dating of overthrust gravels suggest slip rates normal to the fault trace of between 5 and 15 mm a^{-1} (Rattenbury *et al.* 1988, Norris *et al.* 1987), corresponding to total slip rates of 12–36 mm a^{-1}. Strike-slip rates over the last few thousand years estimated by Hull & Berryman (1986) are at least double the dip-slip rates. These slip rates on a 50° dipping fault would give uplift rates at the fault of 4–11 mm a^{-1}, similar to those measured (e.g. Suggate 1968, Adams 1979, Bull & Cooper 1986). Thus a large proportion of the total displacement vector is currently being taken up on the Alpine Fault. Distributed deformation to the east accounts for the balance. Large-scale partitioning of the convergent and fault-parallel components of the plate motion with, for instance, strike-slip being absorbed preferentially on the Alpine Fault and the convergence components distributed to the east (cf. the North Island, Walcott 1978b) does not appear to occur.

The mylonites have formed in schists of the lower amphibolite facies everywhere along the fault from Haast to Arthur's Pass, although within this facies the grade may be somewhat higher in the south due to the greater amount of Mesozoic uplift there (Wallace 1975). These schists were at depths of around 20 km prior to the late Cainozoic uplift, although the maximum depth of burial during the Mesozoic was probably somewhat greater (Cooper 1983). The restriction of the mylonites to the same depth zone and the absence of any deeper crustal rocks exposed along the Alpine Fault despite 70 km of plate convergence suggests that the upper 20–25 km of crust is being delaminated on a ductile detachment and thrust up along the Alpine Fault. The rapid uplift raises the isotherms resulting in high surface heat-flow and a shallow brittle–ductile transition (Allis *et al.* 1979, Koons 1987, Holm *et al.* 1989). The lower part of the crust is presumably imbricated and thickened into the root.

The Southern Alps form a barrier across the prevailing westerly airstream across the Tasman Sea. As a result, rainfall west of the Main Divide is extreme, with values of 10–12 m a^{-1} being recorded. The rates of erosion are correspondingly extremely high, with estimated rates of the same order as the rates of uplift (Adams 1980, Griffiths & McSaveney 1983, Whitehouse 1986). Thus long low-angle overthrusts of the Australian plate by Pacific plate material from the east do not develop as erosion constantly restores the thrust front to the line of the main steep fault. Overthrusts are generally restricted to within a few kilometres of the main fault line (Rattenbury 1986, Norris *et al.* 1987) although some imbrication of the western footwall has occurred with the development of thrust splays within rocks of the western terrane (Rattenbury 1986). A consequence of the rapid erosion of the leading edge of the upthrusting crustal slab is that the topographic, geological and thermal structure west of the Main Divide is essentially in a dynamically maintained steady state, as suggested by Adams (1980) and Koons (1987, 1989).

Deformation east of the Alpine Fault

While major faults may occur parallel to the foliation in the schists, no major repetitions of metamorphic zones are recognized along the West Coast. Further east, however, particularly in northwest Otago, N- to NNE-trending faults are associated with steepening of the schistosity, large-scale folding and repetition of metamorphic and textural zones (Wood 1962, Cooper 1974, Craw 1985). The faults where observable (e.g. Moonlight Fault, Turnbull *et al.* 1975, Craw 1985) are high-angle reverse faults; many may have been rotated from a NW strike by distributed bending (Norris 1979). The imbricate nature of the overall structure indicates shortening of the upper crust above some sort of detachment surface. The orientation of the imbricate faults is approximately perpendicular to the plate vector, so that they are ideally oriented to take up a distributed shortening. The shortening and bending in northwest Otago in part accommodates the northward movement of the Fiordland block (Norris & Carter 1982).

In central and east Otago, the deformation becomes concentrated on two sets of faults, one oriented NE–SW with mainly NW dips and the other striking NW–SE with NE dips, the latter becoming rotated into a more northerly strike towards the west, as discussed in the previous paragraphs. Many of the faults have evidence of Cretaceous normal displacements, although the late Cainozoic displacements on both fault sets is reverse. The two sets are not evenly distributed, with NW-trending faults bounding areas dominated by the NE-trending set (Fig. 1). The two sets of faults appear to have been active simultaneously, both in the Cretaceous and since the late Miocene (Bishop 1974). Late Cainozoic activity on the NW-trending set may slightly pre-date that on NE faults; conversely, most of the evidence for late Quaternary displacement has been reported on NE-trending faults (Makgill & Norris 1983, Beanland *et al.* 1986, Beanland & Barrow-Hurlbert 1988). Total late Cainozoic reverse displacement ranges from a few hundred metres to several kilometres. The total strain decreases eastwards.

Summarizing, the Cainozoic deformation east of the Alpine Fault and, in particular, east of the Main Divide extends as far east as the present coastline, but becomes less intense in this direction. In east Otago and south Canterbury, two sets of almost orthogonal faults, their locations and orientations controlled by pre-existing fault zones, exhibit reverse slip. Further west, the NW-trending faults swing to a more N–S trend, along with greater amounts of reverse slip accompanied by tilting and folding of the foliation and resulting imbrication. This is particularly intense in northwest Otago.

The deformation represents shortening and bending of the upper crust by reverse faulting. To preserve balance, the faults must bottom out into some sort of detachment level deep within the crust. Shortening of the upper crust both along the Alpine Fault and to the east must be accompanied by shortening of the lower crust to form the root. The wider Bouguer gravity anomaly in the south mirrors the wider zone of late Cainozoic imbrication and thrusting in the upper crust. As discussed, most of this shortening is accommodated not on faults parallel to the Alpine Fault but on faults with a more northerly trend. Similarly, the trend of the gravity anomaly is more northerly than the Alpine Fault, as is the zone of deep-seated earthquakes reported by Reyners (1987). Thus, the shortening and root accumulation in northwest Otago is not purely a reflection of the plate convergence normal to the Alpine Fault but also includes a component of plate motion parallel to the fault, consistent with bending and the northward relative movement of Fiordland.

Geodetically measured strain-rates

Walcott (1978a) presented strain determinations from geodetic data in the South Island. The strains were calculated from resurveys of triangulation networks approximately 100 years old. The results show high rates of strain accumulation on the west coast in networks spanning the Alpine Fault, with a principal compression direction trending approximately 115°. This is consistent with the current plate convergence direction and right-lateral reverse slip on the Alpine Fault (Walcott 1979).

Wood & Blick (1986) give details of several years surveillance of fault monitoring networks along the Alpine Fault (Fig. 3). These give comparable orientations to Walcott's data, but with higher strain rates.

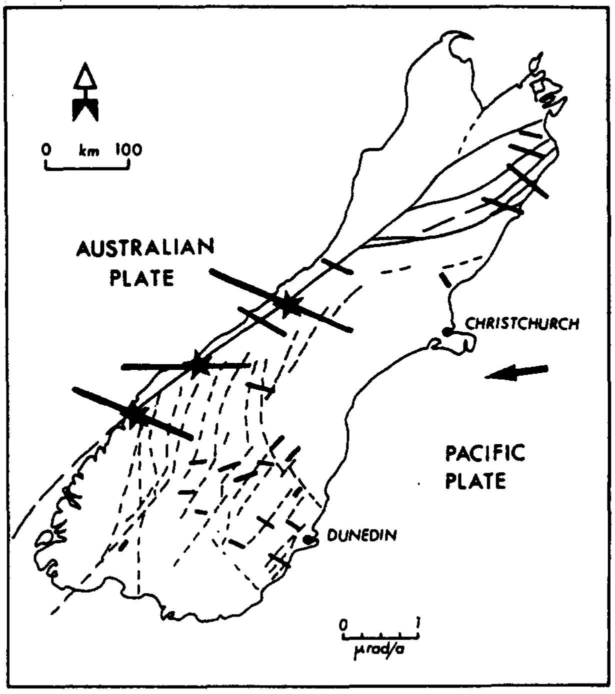

Fig. 3. Map of South Island showing major faults and geodetically measured strain rates. Heavy lines are parallel to the principal compression direction; length of lines proportional to rates. Asterisked measurements are from fault monitoring patterns reported by Wood & Blick (1986) and tend to be higher than regional network data (see text). Remaining data from Bibby (1976), Walcott (1978a), Blick (1986) and Reilly (1986). The heavy arrow represents the direction of motion of the Pacific plate relative to the Australian plate (Walcott 1979).

(These latter are largely the result of high strain rates near the fault being distributed over large first-order networks mainly west of the fault in Walcott's study, compared with the small monitoring patterns across the fault zone employed by Wood & Blick.) Thus the geodetic data are in agreement with the geological evidence reviewed earlier that the strain accumulating in the vicinity of the Alpine Fault represents a substantial component of the total plate displacement vector.

Further east, the situation is more variable. Data from Walcott (1978a), Blick (1986) and Reilly (1986) show that while rates are lower by an order of magnitude compared with the west coast, they are nevertheless significant. The orientation of the principal compression direction varies however from NE to SE (Fig. 3). The compression direction is oriented more northeasterly in the vicinity of the major zone of NW-trending faults, whereas to the south, in areas dominated by NE-trending faults, it is oriented northwest. Towards the west, however, the compression direction swings round, as the fault pattern changes, to a direction closer to that on the Alpine Fault.

The data suggest that significant strain is currently accumulating in areas well to the east of the Alpine Fault, but that it is inhomogeneously distributed and appears to be controlled by the orientation of pre-existing crustal structures. Since the strain on the Alpine Fault is consistent with the present direction of the plate vector, it follows that the integrated deformation further east must also be parallel to the plate vector. It is currently being partitioned differentially due to the inhomogeneity of the crust.

KINEMATIC MODEL

Critical wedge concept

As the component of convergence increased, the plate boundary widened into a zone of continental collision. Recent literature on accretionary prisms and continental collision has demonstrated the usefulness of the concept of critically-deforming wedges in description of the mechanics of collision (e.g. Chapple 1978, Davis *et al.* 1983, Dahlen 1988). In continental collision, a two-sided orogen consisting of two, opposite facing, asymmetrical wedges is produced by convergence of continents with approximately the same thickness. Restriction of the height of indentor to roughly sea level allows the movement of material over the back of the indentor and out of the deforming orogen. In this respect, the collisional orogen differs significantly from that produced by the popular bulldozer analogy of critically deforming wedges (e.g. Davis *et al.* 1983) in which the indentor is assumed to be of infinite height relative to the deforming material. The erosion condition at the indentor may be altered to allow for sedimentation on the indenting plate as in the Himalaya. Such sedimentation results in the development of a fold-thrust belt on the indentor similar to that in the Siwaliks to the south of the Lesser Himalaya.

The two wedges which comprise the collisional orogen are the steep *inboard* wedge facing the indentor which is separated by the main divide from the *outboard* wedge (Fig. 4a) facing towards the undeformed material. In the Southern Alps, the inboard wedge consists of the western draining ranges which lead up to the continental divide, while the outboard wedge is the broad region of deformation which equates to the fold-thrust belt as described by Chapple (1978) in his description of the role of topography in critically-deforming wedges. Although prior to the concept of critical wedges, Wellman's (1979) model incorporated some of these features.

Erosional constraints

The important role of the topographic slope in deformation within collision zones was identified by, among others, Chapple (1978), Stockmal (1983) and Davis *et al.* (1983). Because topography is a function of erosion as well as tectonic uplift, the evolution of a mountain belt can be examined as a coupled mechanical–erosional system. In the Southern Alps, as in many temperate mountain belts, the mountain ranges perturb prevailing wind directions generating a region of heavy rainfall (>10 m a^{-1} on the windward inboard ranges and a rain shadow (≤ 1 m a^{-1}) on the leeward outboard ranges (Griffiths & McSavenney 1983). The asymmetric rainfall

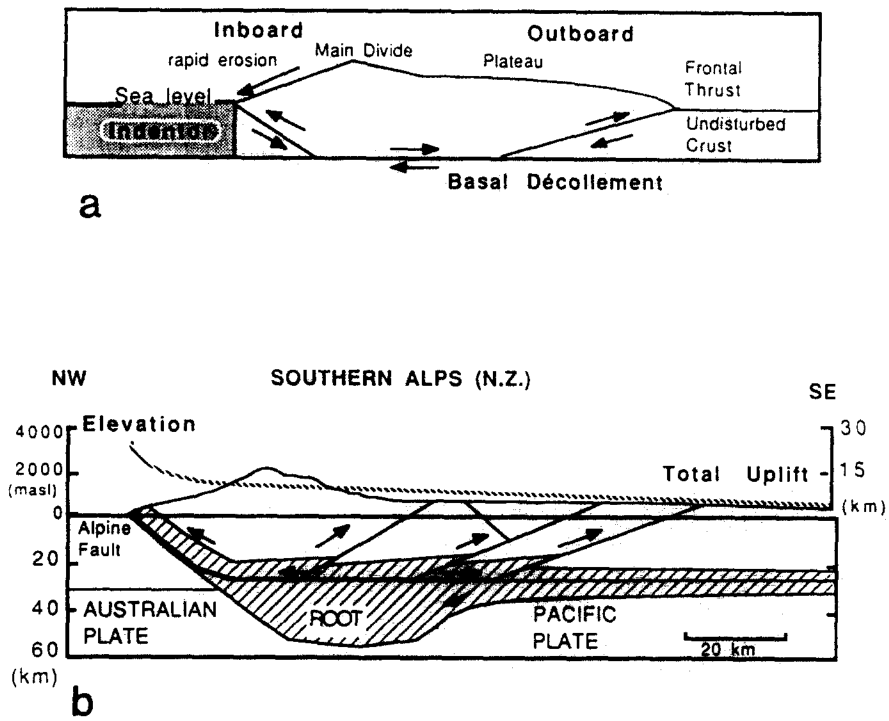

Fig. 4. (a) Schematic representation of two-sided deforming orogen based on critical wedge mechanics with erosional constraints. (b) Schematic section across Southern Alps showing the two-sided character with a lower crust detachment level. Material above this detachment travels through the orogen, undergoing deformation in both outboard and inboard wedges. The high erosion rate of the inboard wedge and the greater strength of the inboard base relative to the basal décollement of the outboard maintains rapid uplift and throughput of material on the inboard side. Material below the detachment accumulates in the root. Cross-hatching represents amphibolite-facies rocks. Hatched curve and scale on right gives distribution of total uplift.

pattern results in a concentration of erosional power in the inboard relative to the outboard. The nature of the combined stream and slope erosion results in further concentration of erosion energy at the toe of the inboard wedge adjacent to the Alpine Fault (Koons 1989). Once material crosses the Alpine Fault, it is rapidly removed from the vicinity of the deforming region by the vigorous ocean currents of the Tasman Sea. In this way the Pacific–Australian boundary remains constrained by erosion to lie within a swath of several kilometres width.

Heavy rainfall on the inboard ranges results in the establishment of near steady-state elevations to the west of the main divide after ~250,000 years of uplift, through coupled stream and slope erosion (Koons 1989). These near steady-state elevations are a linear function of uplift rate, inversely proportional to the erosional transport term and related to the square of the major stream channel spacing. Long-term variation in any of these parameters leads to predictable changes in the near steady-state elevations following a period of transient adjustment from one state to the next. For example, a decrease in the amount of rainfall associated with a glacial epoch results in significant increase in elevation because both the stream spacing and the erosion transport terms are primarily a function of rainfall.

Mountains within the rainshadow are never likely to approach steady-state due to the low rainfall in the region and consequent low erosion constants and high spacings of major streams. Therefore elevations in the outboard region are related to the total uplift and not directly to the uplift rate as in the inboard ranges. For this reason, elevations in the outboard may be high even though uplift rates are low (Fig. 4b).

Eroding two-sided wedge model

If the topographic evolution of a sub-aerial mountain chain described above is imposed upon a deforming orogen in which the top of the indentor remains at sea level, then a rather simple mechanical pattern develops. As convergence increases, the two-sided orogen forms. The basal décollement in the outboard region is a low strength, high-strain zone within the middle to lower crust represented in part by the mylonites currently exposed along the Alpine Fault. The basal surface of the inboard wedge is the Alpine Fault which cuts the entire crust and is significantly stronger than the outboard décollement (Fig. 4b). As compression continues, the orogen grows to the east by thrusting of the outboard wedge over the undeformed material and by forward propagation of the fold-thrust belt, while the present inboard topography becomes established under the very asymmetrical erosion pattern. Relative to the Australian plate indentor, material on the Pacific plate east of the frontal thrust is moving at the full plate velocity. Once incorporated into the orogen by forward propagation of the basal décollement and frontal thrust system, it acquires an eastward velocity relative to the Pacific plate. While its velocity relative to the Australian plate decreases, it remains westward and hence material moves through the orogen and out through the inboard wedge where it is eroded. Strain is significantly concentrated at the toe of the inboard wedge adjacent to the plate boundary (see also Walcott 1978a) as mechanical energy of the deforming orogen is partititioned into the most rapidly eroding part of the orogen. Another high-strain region develops at the outboard toe where material is first incorporated into the deforming wedge. The outboard has thickened by synthetic and antithetic thrusts and now lies in mid-Canterbury and east Otago. The component of plate convergence has increased over time (Walcott 1978a) and therefore the width of the outboard and the height of the inboard ranges have also increased as more material is incorporated into the orogen. Material moves through the orogen with little erosion in the outboard wedge. The width of the near steady-state orogen is primarily a function of erosion in the inboard wedge.

Uplift in the outboard wedge is distributed over a broad region and does not generally exhume deep crustal material. The thermal regime in the outboard wedge is not significantly perturbed by this distributed uplift. Rapid uplift near the Alpine Fault, however, does exhume deep crustal material at the inboard toe. The deep crustal rocks are uplifted at rates too high to allow conductive cooling and a two-dimensional region of high geothermal gradients and of low mechanical strength forms in the upper crust adjacent to the Alpine Fault (Koons 1987). The historical absence of large moment earthquakes in the high-strain region (Reyners 1987) is seen as a consequence of this rapid uplift and therefore ultimately a consequence of long-term meteorological patterns.

IMPLICATIONS FOR THE INTERPRETATION OF ANCIENT BOUNDARIES

Interpretation of slip direction and amount of convergence

The tectonic model for the Southern Alps has many implications for the interpretation of ancient collision zones. The simple two-sided orogen outlined in the last section has two zones of upper crustal deformation above a ductile detachment surface. Material of the colliding plate moves through these zones. The high erosion rate of the inboard wedge maintains the basal western bounding fault essentially in the same location, with very high rates of slip. The outboard wedge on the other hand faces the opposite way and propagates forward more like a classic fold-thrust belt (e.g. Boyer & Elliott 1982). A narrow belt of uplifted rocks from the mid to lower crust is likely to parallel the inboard wedge boundary. After a large amount of convergence, high-grade nappe structures formed at the base of the outboard wedge will be carried through the orogen and be exposed at the inboard side. This has not yet occurred in the Southern Alps.

In the case of oblique convergence the relatively straight, single fault bounding the inboard wedge, which appears as the terrane boundary, may be mistaken for a locus of pure strike-slip displacement and any shortening normal to it calculated erroneously from the thrust displacements in the outboard wedge. The total amount of convergence in the inboard wedge may be very difficult to determine from structural mapping since, unlike the outboard wedge, the overthrust material may be largely eroded away. In the case of the Southern Alps, the only reliable estimates of convergence come from plate motion calculations, yet it is the latter that are usually the goal rather than the source of convergence measurements in old orogens! Estimates of the amount of crustal shortening within the root may be a better method as the root is cumulative with time and therefore is likely to preserve all the evidence, but as it lay within the ductile zone during convergence, problems of exposure and difficulties of analysis will be much greater.

Mylonites uplifted in the hangingwall of the inboard wedge may preserve stretching lineations parallel to the displacement vector. Due to the high rates of slip, however, these lineations will only represent the last increment of displacement. Evidence for earlier increments will have moved through the system and been eroded. On the Alpine Fault, the mylonite fabrics observed at the surface formed at a depth of 8–10 km (Holm *et al.* 1989) and at the present slip rates, would be around 1 Ma old, consistent with young K–Ar dates (Adams 1981). Analysis of the fabrics, therefore, indicates the displacement vector over the last million years or so, when the component of convergence was much greater than in earlier times. The total 480 km of strike-slip on the fault attests to the dominance of strike-slip during the fault's history, but the mylonites do not preserve structural evidence of earlier phases of pure strike-slip movement. Preservation of earlier movement may occur on the footwall side of the boundary. In the northern section of the Alpine Fault, vertical mylonite zones with dextral strike-slip indicators occur entirely within gneisses of the western terrane (Young 1968, Rattenbury 1986, 1987). These mylonite zones are Cainozoic in age and may represent the western side of a broad strike-slip shear zone formed during the earlier history of the Alpine Fault (Rattenbury 1987). Fission track dates of around 10 Ma (White & Green 1986) indicate that these were uplifted and stabilized at about the time the Alpine Fault began to depart from pure strike-slip movement and to develop a component of convergence (Molnar *et al.* 1975, Carter & Norris 1976). These mylonites do not occur everywhere along the length of the fault, however, and where they do, are confined to rocks of the western terrane.

P–T–t *paths within the orogen*

The rapid uplift of the inboard wedge raises material from the depths of the basal detachment faster than it can cool by conduction (Allis *et al.* 1979, Koons 1987, Holm *et al.* 1989). As a result, the thermal structure of

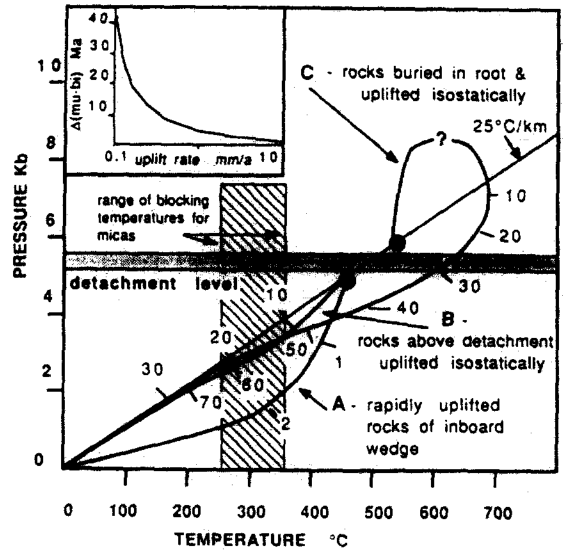

Fig. 5. Diagram showing schematic P–T–t paths of rocks from various sections of the orogen. A is based on Koons (1987) and Holm *et al.* (1989). Others are based on England & Richardson (1977) and England & Thompson (1984). Numbers alongside curves represent Ma after commencement of uplift and for curves B and C are calculated assuming isostatic uplift coupled with erosion, where erosion rate is proportional to relief (see England & Richardson, 1977). Blocking temperatures for micas are based on Hunziker (1979) and Cliff (1985). Inset graph shows variation in difference between muscovite and biotite K–Ar ages as a function of uplift rate (blocking temperature for average muscovite is taken as 350°C and for biotite as 300°C after Hunziker 1979).

the inboard wedge is greatly perturbed, with high thermal gradients near the surface. Thermal gradients within the outboard wedge on the other hand are far less disturbed. While rocks of the inboard wedge are cooling, the rocks of the lower crust imbricated into the root below the detachment are slowly heating up. Immediately upon cessation of uplift, the metamorphic structure at the surface will be characterized by the zone of high-grade metamorphic rocks forming a narrow strip within the inboard wedge adjacent to the boundary. If a large amount of convergence has occurred, this zone will be wider and it will expose nappes repeating the metamorphic sequence. P–T–t paths in these rocks will resemble path A in Fig. 5. (Koons 1987, Holm *et al.* 1989). Due to the rapid uplift, relatively little retrogression is likely except along major shears.

As isostatic uplift of the root occurs, a wider zone of higher-grade rocks will be revealed. These rocks will initially originate above the detachment, so that their history during collision may not involve any substantial pressure or temperature increments. They may, however, exhibit localized zones of high strain. Sense of shear criteria may give contradictory results as the rocks above the detachment are being driven as a whole over the root towards the inboard wedge whereas thrusting within the outboard wedge involves the opposite sense of shear. Their rate of uplift will be considerably slower, by an order of magnitude or more, than the rocks tectonically uplifted in the inboard wedge.

After sufficient uplift to expose the root, a wide zone of high-grade metamorphic rocks will exist on one side of the terrane boundary. Rocks within the root originated in the lower crust, and were then rapidly thickened and slowly heated before slow uplift, so that the *P–T–t* path will be of the type represented by B in Fig. 5. During isostatic uplift of the root, the original crustal detachment will be warped upwards and exposed. Reactivation of the detachment may occur at this time, or subsequently, as a low-angle normal detachment, as has been suggested for thrusts in the Basin and Range Province (e.g. Arabasz & Julander 1986) and elsewhere (e.g. Hossack 1984). Care must be taken, however, in interpreting sense of shear criteria because, as pointed out earlier, different sections of the original detachment may exhibit conflicting senses. Imbrication structures within the root material at an angle to the collision boundary provide evidence for oblique convergence.

Uplift rates and radiometric dating

The *P–T–t* histories outlined above for the various segments of the collision zone will also be reflected in the pattern of radiometric ages. Rocks brought to the surface within the inboard wedge during tectonic convergence will preserve mineral dates of the collision event. Because of the rapid uplift and compression of isotherms close to the surface, only the top 5 km or so are likely to contain rocks below the appropriate blocking temperature at any one time (Koons 1987). In an older orogen, these ages will represent the cessation of convergence. If much uplift has occurred subsequently, this zone of tectonic ages will rapidly be eroded. Rocks below this zone will acquire ages younger than the age of cessation of convergence. With the isostatic uplift of the root, a broader zone will be exposed of rocks with young ages post-dating the collisional event by up to 50 Ma or more (Fig. 5) (England & Richardson 1977, Cliff 1985). Thus it may well be that no rocks preserving ages of the cessation of tectonic convergence will remain after 100 Ma.

One method of testing the origin of the dates appears to be possible. A well-established technique for estimating uplift rates from age determinations is to compare K–Ar or Rb–Sr ages on co-existing muscovite and biotite separates. Since biotite has a lower blocking temperature than muscovite, the difference in ages between the two will depend on the thermal gradient and uplift rate (Fig. 5) (Clark & Jäger 1969, Dewey & Panckhurst 1970, Hunziker, 1979). Tectonic uplift rates in the inboard wedge are relatively high, as is the near surface thermal gradient, so that rocks cooled during tectonic uplift will have virtually indistinguishable muscovite and biotite ages. Rocks having older ages with similar values for muscovite and biotite have been found in some orogens and interpreted as representing tectonic uplift events (e.g. Dempster 1985, Kelly 1988). In contrast, rocks cooled during isostatic uplift will have uplift rates of at least an order of magnitude slower, in addition to lower thermal gradients, so that muscovite–biotite pairs will give measurable age differences indicating uplift rates of around 0.1 mm a^{-1}, typical of many old orogenic belts (e.g. Dewey & Panckhurst 1970, Dempster, 1985). We would suggest that most of the 'typical' uplift rates for orogenic belts determined by this method represent rates of isostatically driven uplift sometime after cessation of convergence rather than actual tectonically driven uplift.

CONCLUSIONS

(1) The Australian–Pacific plate boundary in the South Island of New Zealand is a present-day example of an oblique continental collision zone. Present displacement rates are around 45 mm a^{-1}, with strain partitioned between a narrow, rapidly deforming zone along the western boundary, the Alpine Fault, and a broader zone of distributed deformation to the east.

(2) The collision may be modelled as a two-sided deforming orogen, with the partitioning of deformation being controlled largely by erosional differences between the two sides. The upper crust above a crustal detachment is being rapidly uplifted while the lower crust below the detachment is accumulating as a root.

(3) In ancient collision zones, little evidence may remain of convergence on the inboard side, even though this was the main locus of displacement. Mylonite fabrics are only likely to preserve the last increment of convergence.

(4) Structures in the outboard wedge may be controlled by pre-existing anisotropies and individually are not easily interpreted in terms of plate displacement. The total deformation pattern of the outboard area, if it can be established, may be of more value in identifying the character of the displacement.

(5) Radiometric ages on micas may post-date cessation of collision by tens of millions of years. Dates representing the collision event may not be preserved. Uplift rates calculated from muscovite–biotite pairs will generally reflect isostatically driven uplift. Rocks giving very close muscovite–biotite ages and therefore high uplift rates will relate to tectonically driven uplift.

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