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Geodetic Strain and the Deformational History of the North Island of New Zealand during the Late Cainozoic

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Geodetic strain and the deformational history of the North Island of New Zealand during the late Cainozoic

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The subduction zone under the east coast of the North Island of New Zealand comprises, from east to west, a frontal wedge, a fore-arc basin, uplifted basement forming the arc and the Central Volcanic Region. Reconstructions of the plate boundary zone for the Cainozoic from seafloor spreading data require the fore-arc basin to have rotated through 60° in the last 20 Ma which is confirmed by palaeomagnetic declination studies. Estimates of shear strain from geodetic data show that the fore-arc basin is rotating today and that it is under extension in the direction normal to the trend of the plate boundary zone. The extension is apparently achieved by normal faulting. Estimates of the amount of sediments accreted to the subduction zone exceed the volume of the frontal wedge: underplating by the excess sediments is suggested to be the cause of late Quaternary uplift of the fore-arc basin. Low-temperature–high-pressure metamorphism may therefore be occurring at depth on the east coast and high-temperature–low-pressure metamorphism is probable in the Central Volcanic Region. The North Island of New Zealand is therefore a likely setting for a paired metamorphic belt in the making.

1. Introduction

On the east coast of the North Island of New Zealand the Pacific Plate is being subducted, and immediately to the west is the active rift in continental rocks of the Central Volcanic Region. We have no direct evidence that rocks are being regionally metamorphosed at depth, but there is indirect evidence from surface geology that a substantial accumulation of sediments at high pressure and low temperature may occur in the subduction zone and that silicic magma has been extensively intruded at high levels in the Central Volcanic Region. The North Island therefore provides a likely present-day example of a paired metamorphic belt in the making. The understanding we have developed of the kinematics and tectonic history of the plate boundary zone through New Zealand therefore addresses some of the classic problems of tectonic setting and evolution of such belts.

Today, the boundary between the Pacific and Australian Plates in New Zealand is a zone of deformation 100–400 km wide that, in the south, is dominantly a transform zone and in the north dominantly a subduction zone. That part of the plate boundary zone between Kaikoura on the east coast of the South Island and East Cape on the northeastern corner of the North Island (figure 1) has been variously called the Hikurangi margin (Hatherton 1970), the East Coast deformed belt (Spörli 1978), the East Coast fold and thrust belt, East Coast shear belt, etc. It is a region of Cainozoic accumulation of sediments in an accretionary prism and forms the southernmost part of the Tonga–Kermadec–Hikurangi subduction zone. Parts of it have been active for at least the last 40 Ma although the rate of relative plate motion today, 60 mm a^{-1}, is faster than at any time earlier in the Cainozoic. The tectonics of the region are

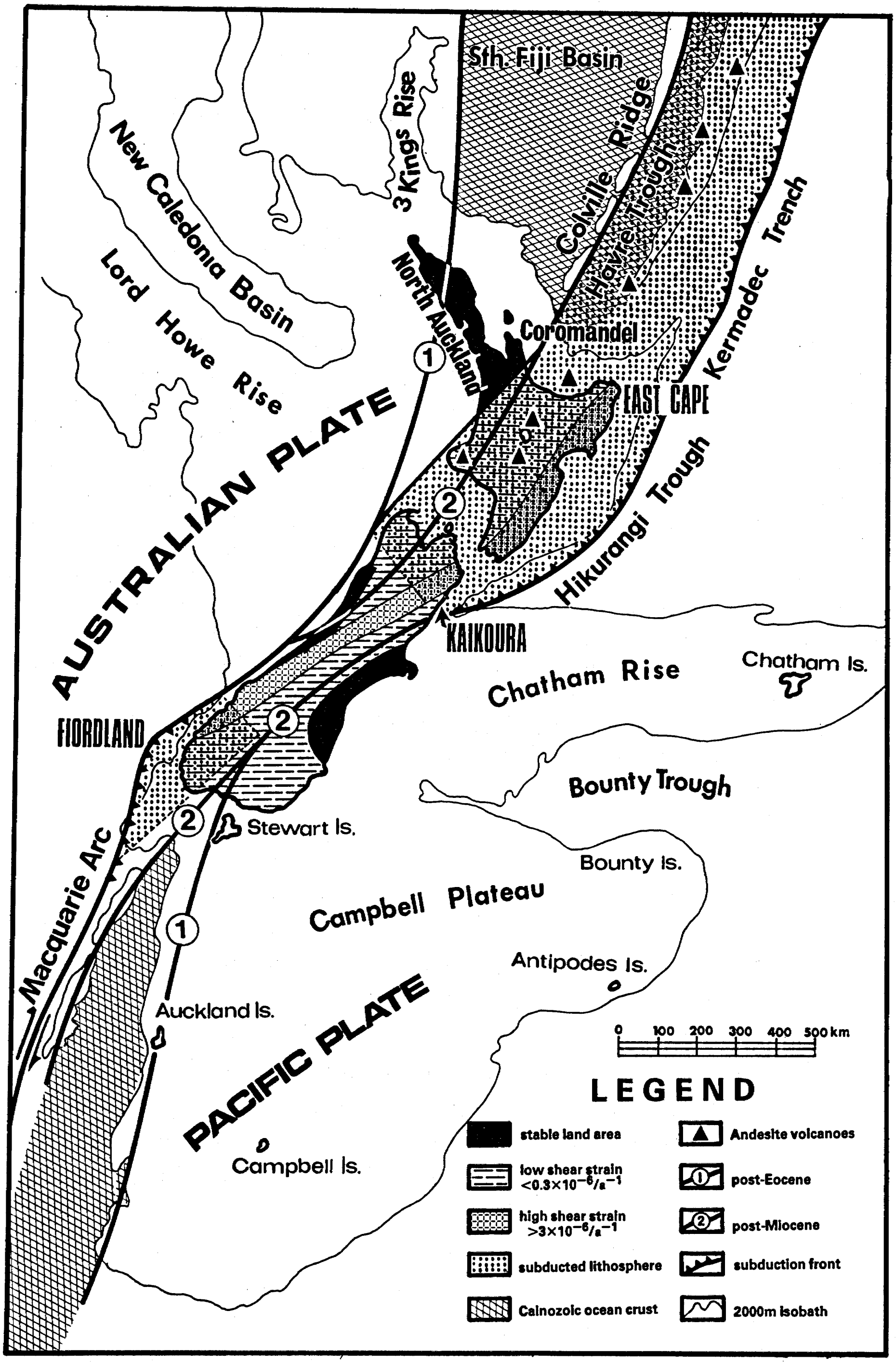

FIGURE 1. The Cainozoic plate boundary zone through the New Zealand region. Cainozoic oceanic crust and subducted oceanic lithosphere lie north and south or New Zealand linked by a belt of high shear strain rate that crosses the country. The ages of oceanic crust and of deformation on land show the location of the Cainozoic plate boundary zone. Line 1 is the outer limit of the post-Eocene and line 2 is the outer limit of the post-Miocene plate boundary zones.

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complex and also involve considerable changes in tectonic style with time, as might be expected at the transition between subduction and transform zones and between oceanic and continental lithosphere.

2. Tectonic setting

(a) *Regional structure*

Apart from two areas to the north and one to the south, the ocean floor around New Zealand is everywhere older than late Cretaceous in age. The two exceptions to the north are the South Fiji Basin of Oligocene age (identified magnetic anomalies range from anomaly 12 (35 Ma) to anomaly 7 (25 Ma)) and the Havre Trough of Pliocene to Recent age (figure 1). To the south, the Emerald Basin contains oceanic magnetic anomalies tentatively identified as anomalies 12 and 13 and, from plate reconstructions, the basin is inferred to have been a region of extension in the period from magnetic anomaly 18 (43 Ma) to anomaly 6 (21 Ma) (Weissel *et al.* 1976).

The outer limit of deformation in the continental crust of the New Zealand region since the beginning of the Pliocene is delineated in figure 1 by line 2 and for the period since the beginning of the Oligocene by line 1. The boundary zone between the Australian and Pacific Plates lay within these limits during the Cainozoic and the areas within lines 1 and 2, including the new ocean floor, are the post-Eocene and post-Miocene plate boundary zones respectively.

The extension on to the earth's surface of the present-day subduction zones identified from intermediate and deep seismicity are shown stippled in figure 1, to the north, east and southwest of New Zealand. Linking the two subduction zones across New Zealand is a belt of high shear strain rate (identified from repeated triangulation surveys) within which the relative motion between the plates is accommodated. The shear strain rate is not uniformly distributed; there is a central part of the belt where the maximum shear strain rate exceeds $0.3 \times 10^{-6}\ \mathrm{a}^{-1}$ (cross-hatched in figure 1) and an outer zone of lower rates of maximum shear strain (lined).

(b) *Geodetic strain*

The current deformation of the plate boundary zone through New Zealand on a time scale of a few decades is shown by the two shear-strain rate components derived from repeated triangulation surveys and the velocities calculated from them (figure 2). The differences in the observed angles between first and subsequent survey(s) are directly related to the shear strains γ_1 and γ_2 that are defined by $\gamma_1 = \partial u/\partial x - \partial v/\partial y$, $\gamma_2 = \partial u/\partial y + \partial v/\partial x$ where u and v are the displacements in the direction x and y respectively. In figure 2*a* the x axis is taken to be parallel to the local trend of plate boundary zone, which differs in North and South Island. The $\dot{\gamma}_1$ component is the shear strain *rate* that corresponds to a pure shear involving shortening (positive γ_1) or extension (negative γ_1) along the y axis balanced by thickening or thinning in the vertical direction. The $\dot{\gamma}_2$ component corresponds to a simple shear strain rate which is positive for dextral shear in the x direction.

The particular value of geodetic strain to the study of the tectonics of a region is that it gives a direct measure of the rate of deformation as well as the orientation of the strain ellipse. Whereas the latter can also be inferred from fault-plane solutions of earthquakes and such geological observations as fault slip and fold orientation, only geodetic strain gives the shear strain rate. Where it can be compared with other data it agrees very well with them. The integrated shear strain rates across the South Island between the Pacific and Australian Plates

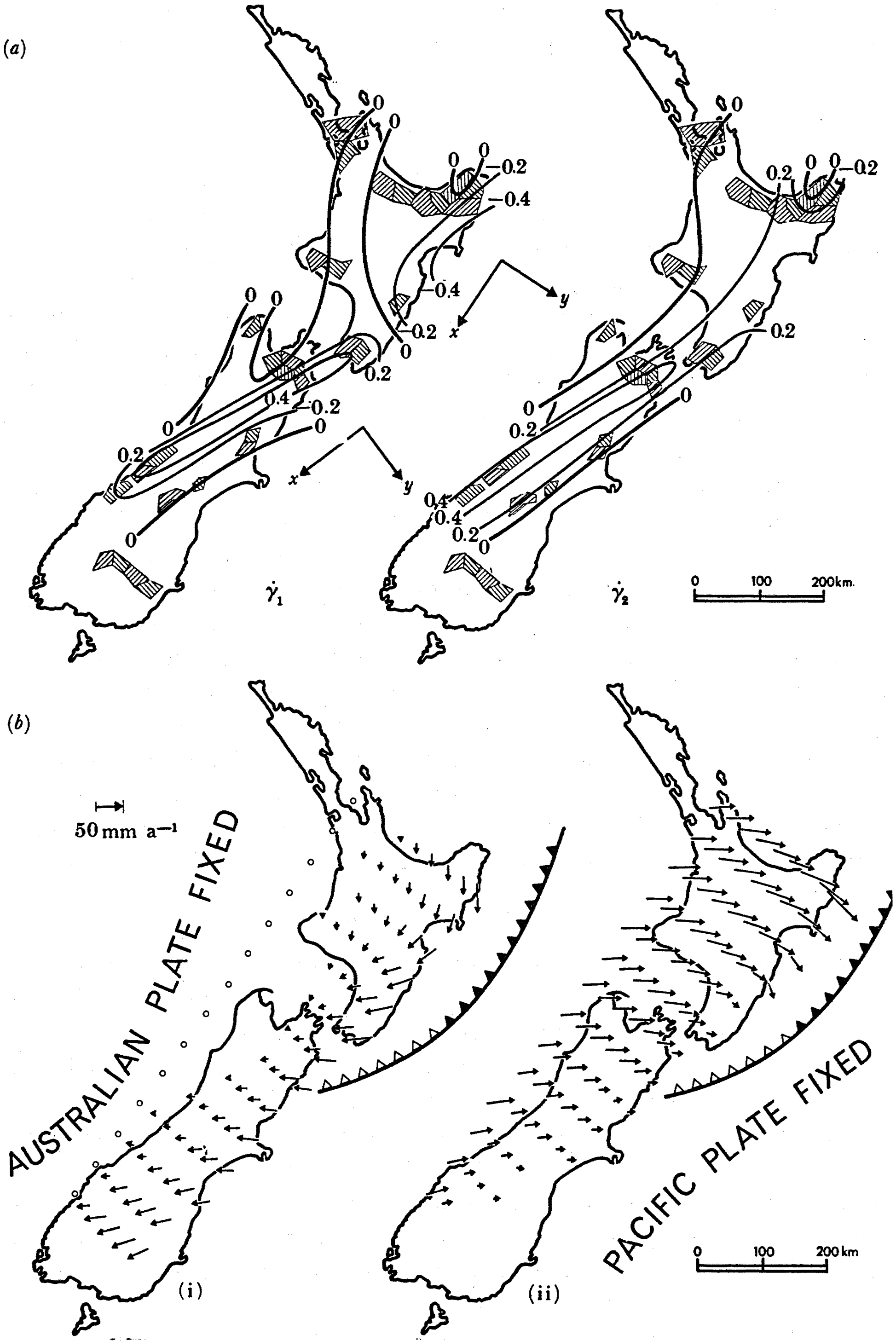

FIGURE 2. (*a*) Rates of shear strain (units of 10^{-6} rad per year) for the components $\dot{\gamma}_1 = (\partial u/\partial x - \partial v/\partial y)$ and $\dot{\gamma}_2 = (\partial u/\partial y + \partial v/\partial x)$, with axes as shown, differing in North and South Islands. The contours are based on the average values estimated for each of the lined regions. (*b*) Velocities derived from shear strains. Both from Walcott 1984*b*.

give the same direction and speed of the two plates as those estimated from sea floor spreading (Bibby 1981). Where the estimates of geodetic strain are derived from triangulations that cover large areas (the data for figure 2 come mostly from repeated first-order triangulation quadrilaterals that measure 40×40 km) the irregularities introduced by earthquakes smaller than about magnitude 7.5 are smoothed out (Walcott 1984*b*). Thus maps of geodetic strain or the velocity maps derived from geodetic strain illustrate the tectonic deformation on a scale comparable to crustal thickness and to the major tectonic elements of a region.

(*c*) *Velocities*

Estimated velocities at regularly spaced points within the plate boundary zone are shown in figure 2*b*. They are calculated by integration of the two shear strain rate components using the method of Haines (1982). The two patterns are kinematically equivalent, figure 2*b* (i) being the pattern obtained by holding the Australian Plate fixed and figure 2*b* (ii) obtained by holding the Pacific Plate fixed; the transformation from one to the other is effected by the Chase (1978) Euler rotation.

The velocity maps are useful in that they show very concisely the large scale kinematics of the plate boundary zone. As is to be expected, the kinematic pattern over the North Island is dominated by the subduction zone. Relative to the Pacific Plate, the direction of motion throughout the Hikurangi margin is orthogonal to its trend; the component of the oblique motion of the plates parallel to the trend is taken up by strike-slip on the transcurrent faults (figure 3) behind the Hikurangi margin. In the southern half of the Hikurangi margin all of the relative motion of the plates is accommodated within the overlying crust, i.e. there is no slip on the subduction zone and the décollement can be said to be locked (Walcott 1978*b*).

(*d*) *Structure of the Hikurangi margin*

The change from compression to extension in the direction perpendicular to the trend of the plate boundary on the east coast of the North Island occurs near a line (figure 2*a*), that crosses the coast at Cape Turnagain (figure 3) and there are considerable differences in Cainozoic structure and geology north and south of it. South of the line, late Cainozoic sedimentary rocks occur in flysch basins a few tens of kilometres in length separated by antiformal basement highs of Cretaceous and older rocks thrust along west dipping faults (van der Lingen 1982). The structural association is a fold and thrust belt and it extends from the Hikurangi Trough westward to the transcurrent fault zone and the axial greywacke ranges.

North of the line the fold and thrust belt angles offshore and is separated from the axial ranges by a large fore-arc basin, the Wairoa Basin, 250 km long and 80 km at its widest. North of Hawke Bay the basin contains an essentially continuous sequence from Cretaceous to Pleistocene. The Cainozoic sedimentary rocks older than mid-Miocene are derived from fine grained, bentonitic clays with tuffaceous bands and indicate slow sedimentation rates in rather deep water. About 20 Ma ago, rapid sedimentation of alternating siltstones and mudstones began and continued into the Pliocene with some 4 km accumulated in the period 15–5 Ma (Kingma 1964). West and southwest of Hawke Bay a sequence of Plio-Pleistocene sedimentary rocks 3 km thick accumulated in a fore-arc basin in which the sedimentary rocks onlap the basement in the late Miocene along the axial range to the west, and offlap during the Pliocene.

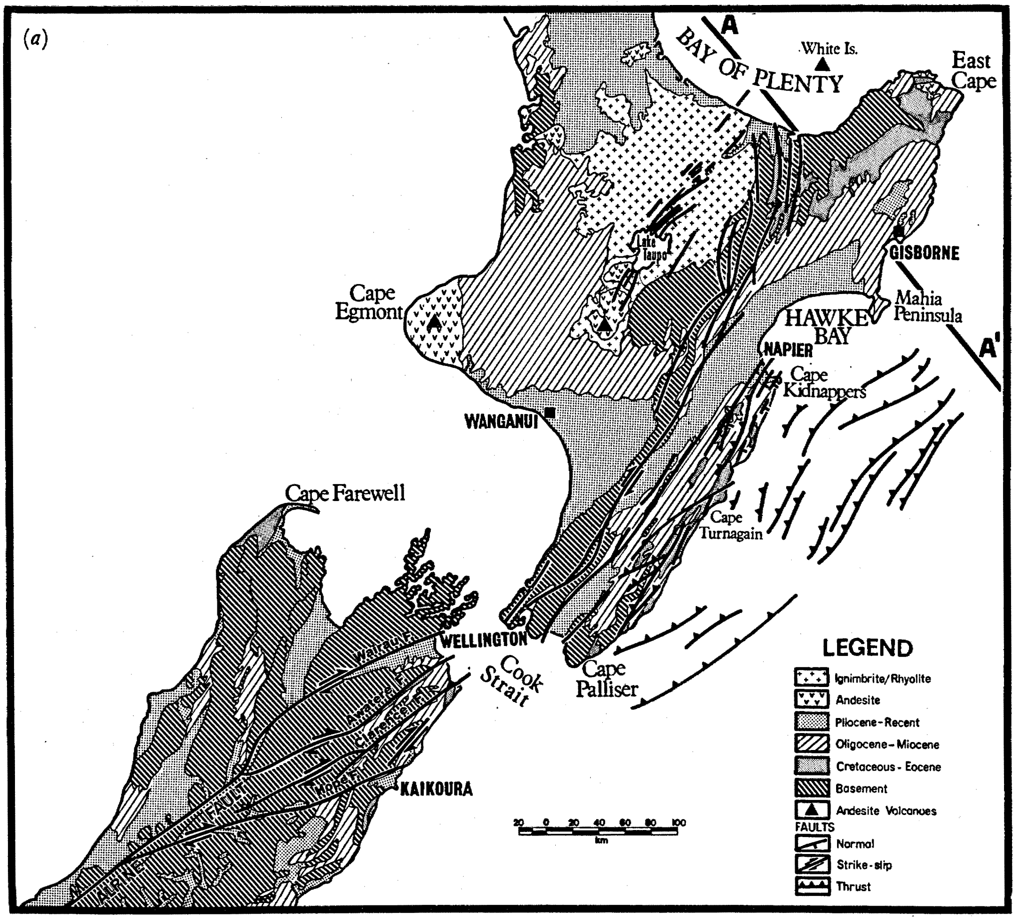

FIGURE 3*a*. Tectonic map of the Hikurangi margin. A–A′, line of section of figure 9.

(*e*) *The Central Volcanic Region*

West of the Hikurangi margin is the Central Volcanic Region, a triangular area with apex at Ruapehu and base along the Bay of Plenty coast (figure 3). Its principal structural features are summarized by Stern (1985) who, like Karig (1970) and others, argues that it is a structural continuation of the Lau and Havre Trough back-arc spreading into the continental crust of New Zealand. It is an active region of extension, intensive volcanism, very high heat flow and thin crust. The basement, believed to be largely of Mesozoic greywacke, has subsided about 2 km within the last 1 Ma, but the more than 10^4 km^3 of largely acidic volcanics erupted in the same period (Cole 1981) has filled, and even raised a volcanic plateau above, the subsiding basin. Volcanic activity today is confined to the eastern third of the Central Volcanic Region, the Taupo Volcanic Zone where heat flow averages 700 mW m^{-2} or about 12 times the continental average. A reversed seismic refraction profile indicates that a layer of compressional velocity 6.0 km s^{-1} and thickness 15 km overlies a 7.4 km s^{-1} layer (Stern 1985) and the upper mantle velocity (Pn) from earthquake travel times also gives about 7.4 km s^{-1} for the Central Volcanic Region (Haines 1979).

Shear strain rates estimated from repeated geodetic triangulation surveys near the Bay of Plenty (figure 2*a*) average about 0.1×10^{-6} radians a^{-1} over a distance of 120 km in both $\dot{\gamma}_1$

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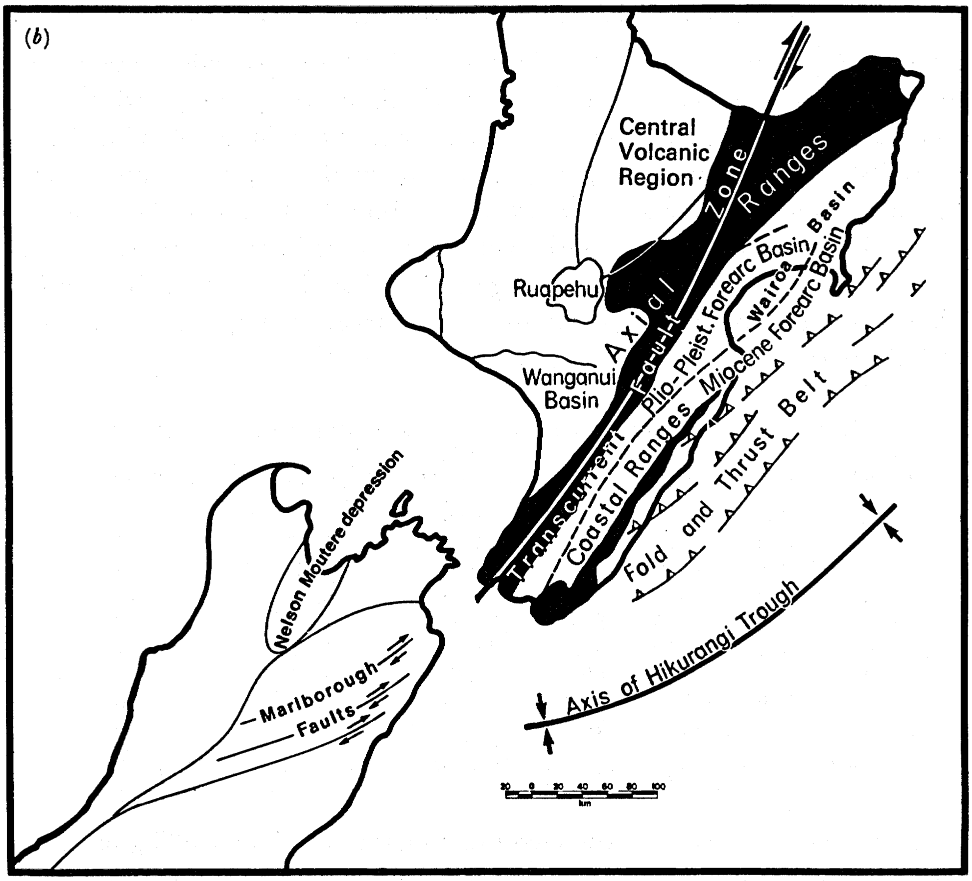

FIGURE 3*b*. Names of major localities and settings.

and $\dot{\gamma}_2$. Integrating over the region gives a velocity of about 12 mm a^{-1} for extension and the same for dextral shear. Repeated triangulation surveys of lesser accuracy indicate a similar orientation of the strain ellipse throughout the Central Volcanic Region (Walcott 1978*a*; Sissons 1980).

The coefficient of extension, the β value, indicated by the present crustal thickness of 15 km is about 2, assuming an original crustal thickness of 30 km. However, it could be greater if a substantial part of the thinned crust were made up of cooling introduced igneous rock as suggested by Stern (1985) to account for the exceedingly high heat flow.

The volcanic rocks of the Central Volcanic Region are very young, less than 1 Ma old, although the extension itself has probably occurred over a much longer period: the present day rate (12 mm a^{-1}) near the Bay of Plenty coast would produce the required extension in about 5 Ma. This is similar to the age of spreading of the Havre Trough.

Older volcanic rocks and centres of activity are exposed to the northeast in the Coromandel and North Auckland peninsulas. The oldest rocks are about 20 Ma in age. Thick accumulation of sediment derived from acidic volcanic rocks in the Miocene and Pliocene along the Hikurangi margin demonstrates that intensive volcanic activity has had a much longer history than that of the Central Volcanic Region.

(*f*) *Episodic changes in the γ_1 component of shear strain*

In the central North Island and all of the South Island, the geodetically determined shear strain rates do not appear to have changed with time. Bibby (1975) compared strain rates from surveys that cover different periods in the northern part of the South Island and showed that the two shear strain rate components were constant in the last 100 years. Moreover, the integrated shear strain rates across the South Island from the Pacific to the Australian Plates give the same direction and speed for the two plates as that estimated from sea floor spreading, namely the Chase (1978) Euler parameters given in table 1. Thus over the South Island of New Zealand the shear strain rates determined from survey data appear to have a long-term stability (Walcott 1984*b*).

Table 1. Pacific/Australian finite rotation parameters

anomaly	age/Ma	lat.	long.	angle/deg	rate/deg Ma^{-1}	rate†/mm a^{-1}
5	9.8	−56.4	184.8	12.13	1.24	39
6	19.5	−52.5	181.1	20.04	1.03	24
13	35.6	−51.6	181.0	37.76	1.06	23
13a	35.6	−49.2	177.4	33.26	0.93	15
13b	35.6	−50.7	183.0	36.20	1.02	22
13c	35.6	−53.8	185.0	43.39	1.22	34
13d	35.6	−52.4	180.0	38.76	1.09	25
18	43.0	−50.5	179.8	44.15	1.02	19
present	0	−62.0	−174.3	—	1.27	50

† Rate of relative plate motion at 'Wellington', lat. −41.2°, long. 174.8°.

This is not, however, the case for the Hikurangi margin north of Wellington. Currently, in the northeastern part of the North Island, $\dot{\gamma}_1$ is negative indicating extension perpendicular to the plate boundary. Before the magnitude 7.9 Hawke Bay earthquake of 1931, $\dot{\gamma}_1$ was positive indicating shortening perpendicular to the plate boundary zone at that time (Walcott 1978*b*). It has been suggested that the long-term trend is compressional, with the present extension being the relaxation of previously imposed shortening (Walcott 1978*b*, 1984*b*). However, there are several reasons, given below, to suggest that in the northern part of the Hikurangi margin the opposite is true; the long term trend in the fore-arc basin is extensional and this is interrupted from time to time by short periods of compression such as that immediately prior to the 1931 Hawke Bay earthquake.

(*g*) *Extension on the Hikurangi margin*

The general structure of the Wairoa Basin is synclinal with a hinge along the western Hawke Bay coastline extending northeastwards toward East Cape. The vertical deformation associated with the 1931 earthquake involved uplift of the hinge of the syncline, which is the opposite of the long term trend. The southeastern limb of the syncline is extensively cut by normal faults that dip commonly to the southeast, i.e. towards the Hikurangi Trough. The significance of the normal faulting is, however, open to differing tectonic interpretations. Normal faulting parallel to the coastline south of Cape Kidnappers and extending over a distance of 20 km has been interpreted as the head of a southeast-directed slump (Pettinga 1982). Closely spaced, active normal faults extending over a width of 5 km at Cape Kidnappers are inferred to be secondary features developed over an anticlinal arch above a presumed active thrust at depth (Hull 1985).

Normal faulting is more extensive than these two interpretations in terms of local structure would suggest. A major normal fault is described from seismic reflection profiles on land south of Napier (O'Halloran 1963). About 80 km southwest of East Cape extensively developed normal faults are well described by Mazengarb (1984). The normal faults are spaced at an average width of about 6 km and dip to the southeast at about 30°. A displacement on one fault of about 2–4 km is inferred. The sedimentary rocks between the faults dip northwest at about 20–30°, indicating a strain of about 1.4 if the strain is expressed in the tilting of the blocks.

Extension by normal faulting in the fore-arc basin therefore appears to be of regional and not merely local significance and, like other regions of extension, has presumably resulted in tilting of the strata and rotation of the normal faults from initially steep to subsequent gentle dips: this is the interpretation given in figure 9.

3. Tectonic history

(a) *Relative plate motion*

The relative plate motions during the Cainozoic in the southwest Pacific are among the best determined anywhere, because of a good knowledge of the transform faults and magnetic anomalies along the Antarctic Plate boundary and a favourable geometry of the three plate system: Australia, Antarctica and Pacific Plates. The most recent analysis of these data is that of Stock & Molnar (1982) who list rotation parameters that include 'best-fit' and 'range' of possible values for each of the Antarctic–Pacific and Antarctic–Australian boundaries and for each of finite rotations that bring anomalies 5, 6, 13 and 18 into coincidence from either side of the Antarctic Plate boundary.

Stock & Molnar (1982) have pointed out that the 'best-fit' rotation poles for anomalies 6, 13 and 18 lie within the range of the position for the rotation pole of anomaly 13 and, hence, during the period from 43 to 21 Ma ago the instantaneous pole position is likely to have changed hardly, if at all. Neither did the rate of rotation, as the average is much the same for all three finite rotations (table 1). The rotation parameters for anomaly 5 lie well outside the range of anomaly 13 and it is likely that the instantaneous pole began moving away from New Zealand between 10 and 21 Ma ago. The increase in distance was accompanied by a faster rate of rotation and the rate of relative plate movement increased dramatically along the plate boundary in the vicinity of New Zealand (table 1). The 'instantaneous', as distinct from the average, rate of motion near Wellington increased from less than 20 mm a^{-1} between 15 and 45 Ma ago to about 50 mm a^{-1} today. The position of the pole for the Euler rotation (Chase 1978; Minster & Jordan 1978) differs considerably from that of the finite rotation pole of anomaly 5, so there has been a change in direction of about 20–30° as well as in rate of motion between the plates in the vicinity of New Zealand.

The total displacement across the post-Eocene plate boundary zone is about 900 km (Stock & Molnar 1982) and the total displacement across the post-Miocene plate boundary zone is about 200 km, obtained from the average rate in the last 5 Ma.

(b) *Reconstruction of the plate positions*

By using the finite rotations derived from Stock & Molnar (1982), it is a straightforward procedure to derive the relative position of the plates at the times of identified magnetic anomalies. However, instead of obtaining reconstructions for these irregular periods of time the rotation parameters have been determined at 5 Ma intervals by interpolation (figure 4).

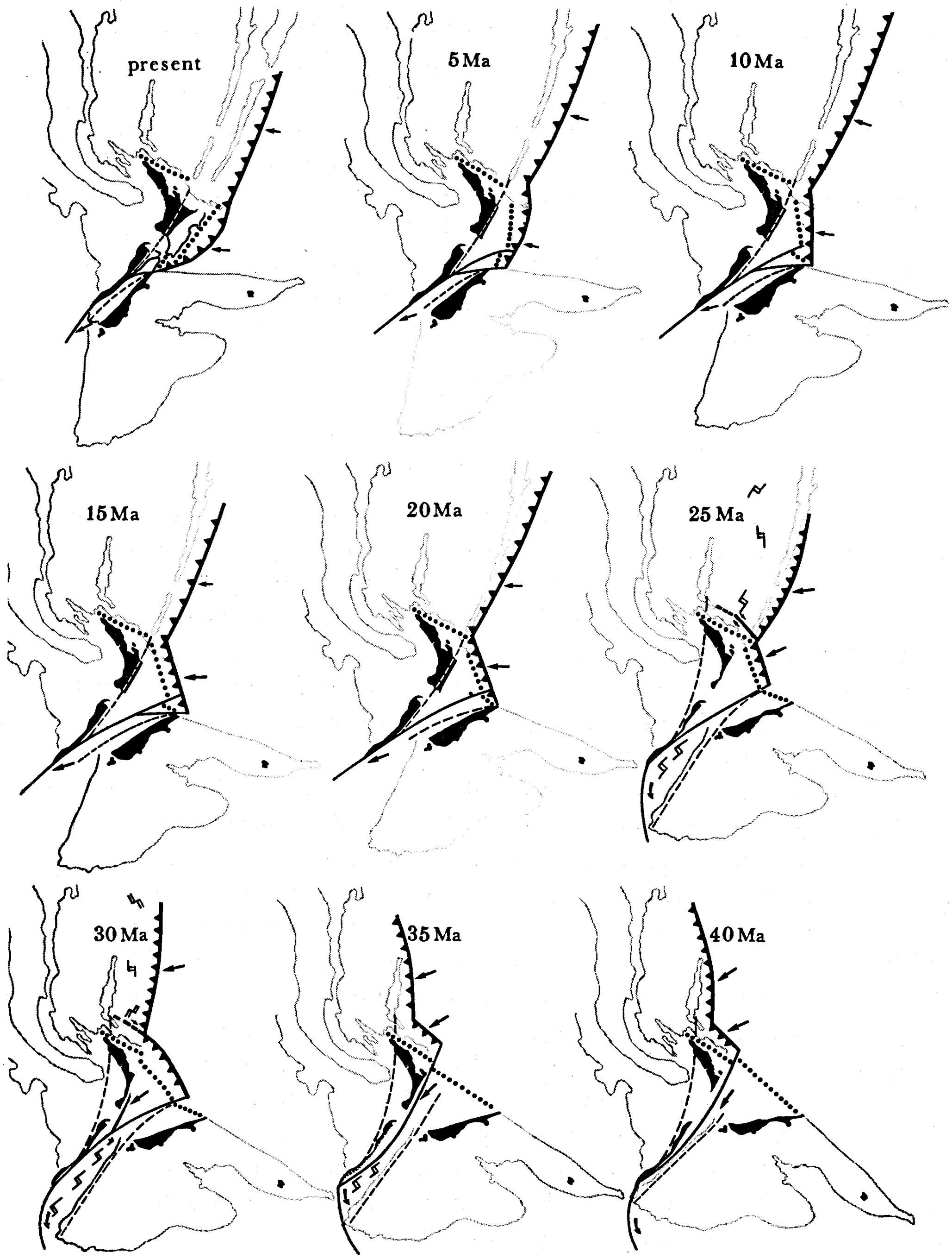

FIGURE 4. Reconstruction of the plate boundary zone for the New Zealand region. A reference line (dotted) at the continent–oceanic boundary displays the deformation within the plate boundary zone. The broken lines correspond to 1 and 2 of figure 1. The position of the major subduction zone north of New Zealand is determined by the age of oceanic crust and an assumption that there has been a single west-facing subduction zone in the last 40 Ma. Reconstruction is made by assuming the trench to bear the same spatial relation to the extinct arc of the Colville Ridge (5–25 Ma) and the Three Kings Rise (35–40 Ma) as the present trench does to the present arc. The position and orientation of the linking (Hikurangi) subduction zone and the transform (Alpine Fault) are constrained by structures in the central part of the South Island, the palaeomagnetic data of figures 7 and the relative plate positions.

The estimate for both position of the pole and angle of rotation of the 5 Ma reconstruction may be in error because of the rapidly changing values of the rotation parameters in the late 10 Ma, but the pole positions for the other reconstructions do not differ greatly from the original and only the rotation angle is affected by the interpolation.

The reconstructions of the plate boundary zone show the inferred position of the trench related to the subduction zone north of New Zealand; the trench is drawn in the same relative position to the volcanic arc as it is today in the Kermadec Arc. The position of the volcanic arc for the period 25–5 Ma ago is taken to be the Colville Arc and for the period 40 and 35 Ma ago is taken to be the Three Kings Rise. Neither assumption is supported by any dated materials from these areas, and they are based on the supposition that both the South Fiji and Havre oceanic crusts were formed by back-arc spreading behind a long-lived Pacific–Australian subduction zone that has migrated persistently eastward.

In figure 5 the trenches at successive periods are superposed on a 40 Ma reconstruction and illustrate the progressive subduction of the Pacific oceanic lithosphere. The heavy broken lines mark the position where the subducted plate changed direction substantially (as it does, for example, today near Honshu) and presumably mark major tears in the plate where one subducted limb overrides the other. The dotted line is the limit of present day seismicity in the subducted plate: we can infer that a large part of the subducted oceanic lithosphere has been resorbed and is no longer seismically active.

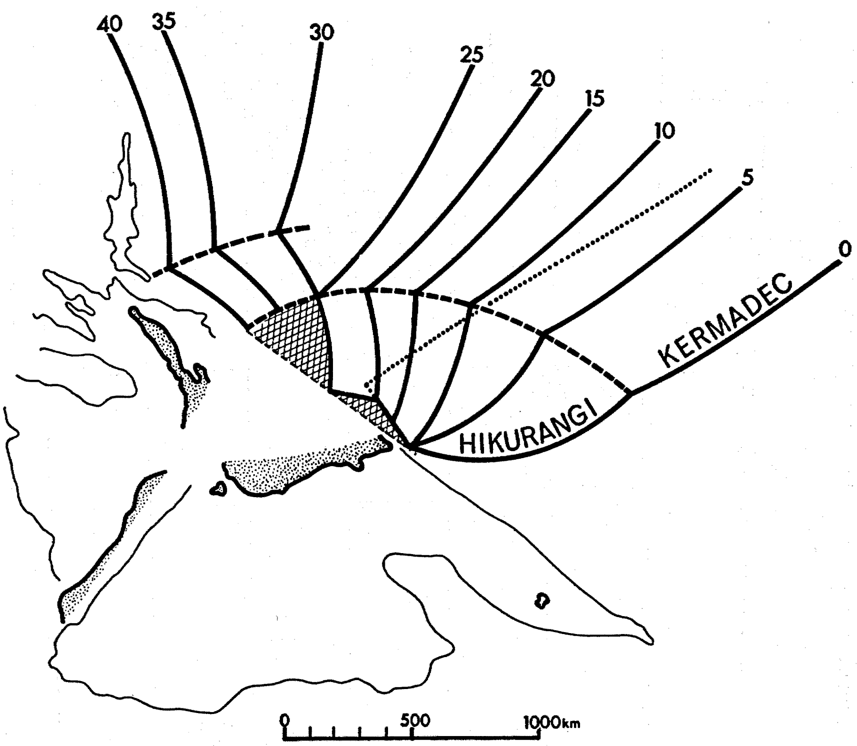

FIGURE 5. The successive positions of the trenches marking the subduction zones in figure 4 plotted on the 40 Ma reconstruction of the plates. Note how the Pacific Plate has been progressively subducted with particularly rapid subduction occurring in the periods of back-arc spreading (35–25, 5–0 Ma). The dotted line marks the area of the presently seismically active subducted plate. Most of the inferred subducted plate is aseismic. The broken lines show the inferred position of tears in the subducted plate.

The reconstructions of figures 4 and 5 show a clear distinction between the Hikurangi and Kermadec subduction zones. They are separated by a tear in the Pacific Plate with the Hikurangi but not the Kermadec zone having rotated very considerably during its history. More than 1000 km of lithosphere has been subducted in northern parts of the Hikurangi margin but little or none has been subducted in the south; the Hikurangi margin has swung like a beam on hinges that lies near Kaikoura and East Cape.

The southeastern limit of the Hikurangi subduction zone has moved from a position near Coromandel 35 Ma ago, to near Cook Strait 25 Ma ago and reached its present position near Kaikoura between 10 and 15 Ma ago. Whether the southeastward propagation occurred at discrete times as shown in the reconstructions (which would result in the non-subduction of the checked areas (figure 5)) or occurred progressively (where these areas could have been subducted) is uncertain. Progressive motion of the end of the subduction zone requires a progressive eastward migration of the transform system through the continental crust. Something like this appears to have occurred in the last 20 Ma with the development of the

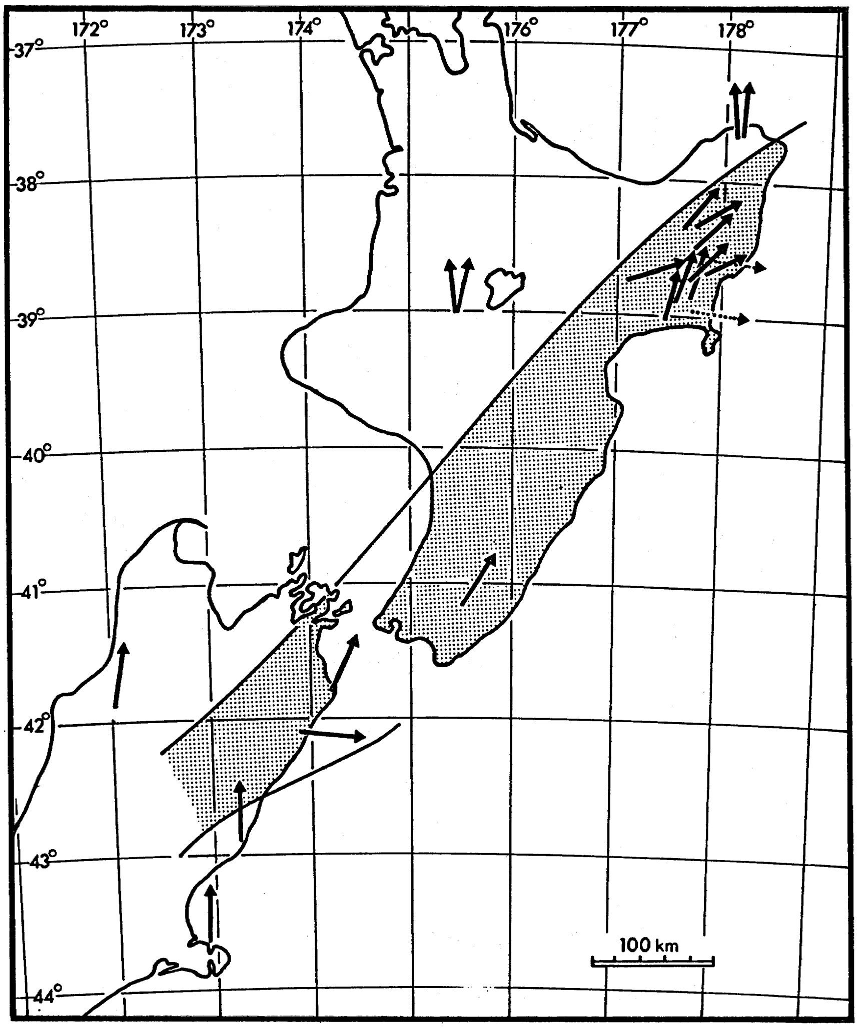

FIGURE 6. The declination of primary magnetization in sedimentary rocks less than 25 Ma old in Central New Zealand. The expected direction without rotation about axes is north–south. The shaded area is the zone of large rotations and corresponds to the zone of present day deformation (figure 2).

Marlborough Fault system in northeastern South Island (figure 3); the most southerly fault appears to be taking up most of the relative plate motion today, whereas the Wairau Fault to the north took up most of the early motion.

(*c*) *Rotation of the Hikurangi margin*

Reconstruction of the plate boundary zone through New Zealand requires substantial changes to have occurred in the orientation of the Hikurangi margin during the Cainozoic. Palaeomagnetic studies over the last few years (Walcott *et al.* 1981); Walcott & Mumme 1982; Mumme & Walcott 1985) show that such rotations have indeed occurred and prove particularly important in both determining the geographic extent of the deforming region and constraining the reconstruction models. Moreover, the rotation is occurring today as shown by the geodetic strain analysis.

All sites for which reliable palaeomagnetic data are available on the Hikurangi margin south of East Cape and north of Kaikoura show the direction of primary magnetization to have been rotated around vertical axes (figure 6) with the amount of rotation increasing with the age of the rocks. A summary of data from the Wairoa Basin is given in figure 7. The three largest rotations (reported in Walcott & Mumme 1982) are from sites that had suffered additional local deformation, as their strike differs from the regional value.

The rate of present day rotation can be determined directly from the velocity maps in figure 2*b*. Near East Cape the velocity of the Hikurangi margin relative to the Pacific Plate is about 75 mm a^{-1} directed perpendicular to the trend of the plate boundary zone. Southwestwards along the coast this direction remains constant but the rate decreases steadily to zero at Kaikoura, 600 km from East Cape. The rate of rotation is therefore 7° Ma^{-1}.

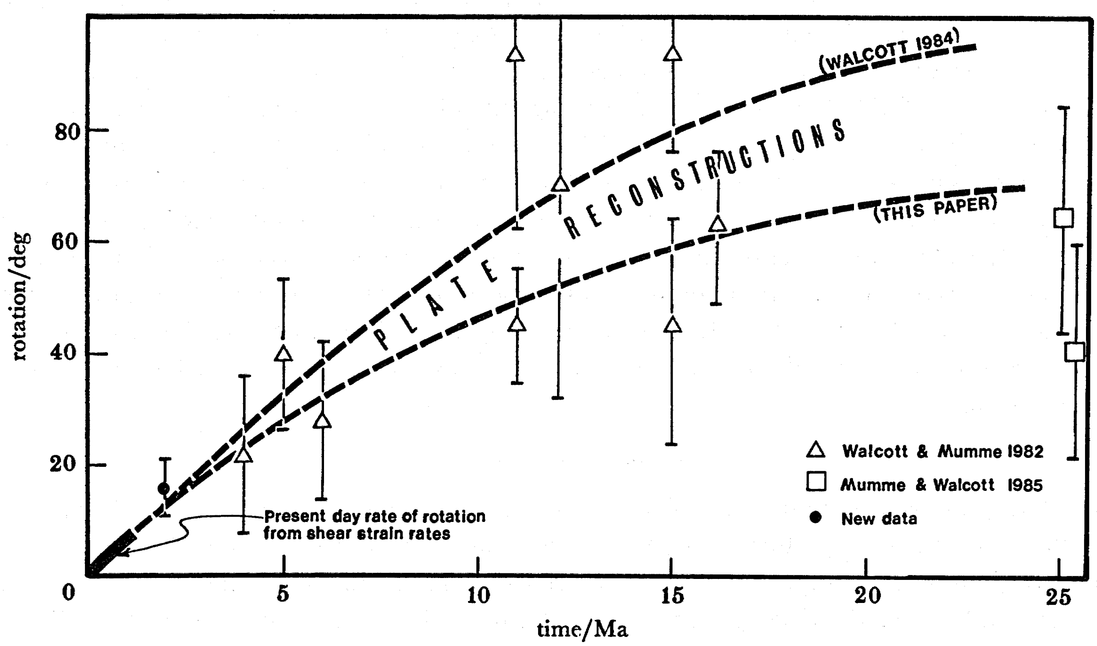

FIGURE 7. Rotation of the Hikurangi margin with age. Plotted data show mean value and 95% confidence limits on anomalies in declination for sites within the Wairoa Basin. The broken lines show the rotation of the Hikurangi margin implied by the reconstruction of Walcott (1984*a*) and this paper (figure 4).

Earlier interpretations presumed there to be a linear change in declination with time, with a rate of about 3° Ma^{-1}. However, this conflicted with the present-day rate (Walcott 1984*b*). Recent palaeomagnetic studies of sedimentary rocks 2 Ma old (details of which will be reported elsewhere) give a declination of about 16° and confirm that the rate of rotation has increased toward the present.

The rotations from the reconstruction models (figure 4) are derived from the orientations of the dotted line representing the Hikurangi margin. An earlier model given in Walcott (1984*a*) required considerably greater rotation than that of the preferred model given in this paper.

Palaeomagnetic studies at East Cape and south of Kaikoura show that no rotation has occurred in these regions. The boundaries between the rotating and non-rotating parts of the Hikurangi margin are constrained to within 50 km. The nature of the structures developed in these 'hinge' regions is uncertain and is the subject of present study.

4. Accumulation of sediments to the Hikurangi margin

(*a*) *Accretion of sediments to the frontal wedge*

The thickness of sediments above the basement reflector in the Hikurangi Trough immediately in front of the fold and thrust belt is about 3 km opposite Hawke Bay (Lewis 1985, figure 13), and post-Miocene sediments alone are more than 3.5 km thick opposite Cook Strait (Katz 1982, figure 5). The present topography and, presumably, the rate of erosion are greater today than earlier in the Cainozoic, so the rate of sedimentation is also likely to be greater than in the past. I take the thickness of sediments in the Hikurangi Trough in the early Miocene (25–15 Ma) to be 1 km, in the late Miocene (15–5 Ma) to be 2 km and in the last 5 Ma to be 3 km. The amount of subduction in the vicinity of Hawke Bay can be estimated from figure 5 as 300 km in the early Miocene (30–20 Ma), 400 km in the late Miocene (20–5 Ma) and 300 km in post-Miocene times. These values together give a volume per km in length of the subduction zone of 300, 800 and 900 km^3 during the 3 periods, or a total between 1000 and 3000 km^3 as the uncertainty in sediment thickness must be about 50%. The volume of the frontal wedge southeast of the coastline and the fore-arc basin is about 1000 km^3 (18 km thick at the coastline, based on the depth of the hypocentre of the 1966 Gisborne earthquake (Webb *et al.* 1985), 120 km from the trough); yet a substantial part is of rocks older than Miocene as Cretaceous and Eocene outcrop on the sea floor well towards the trough, and surface rocks are of slope rather than trench derivation (Lewis 1985). It is likely therefore that a substantial part of the sediment accreted into the subduction zone has been transported below and beyond the fore-arc basin.

(*b*) *Late Quaternary uplift*

The late Quaternary uplift of the Hikurangi margin is shown in figure 8, derived from maps by Pillans (1986) and Wellman (1979). Apart from the two regions shown as insets that demonstrate the detail and complexity of the deformation, the data on which the map is based are sparse and scattered. Observed rates exceed 3 mm a^{-1} but these tend to be localized. There is a broad belt extending southwestward of East Cape along the axial ranges where the rate of uplift in late Quaternary times exceeded 1 mm a^{-1}. The belt coincides with the negative gravity anomaly related to the subduction and both are parallel to the strike of the subducted

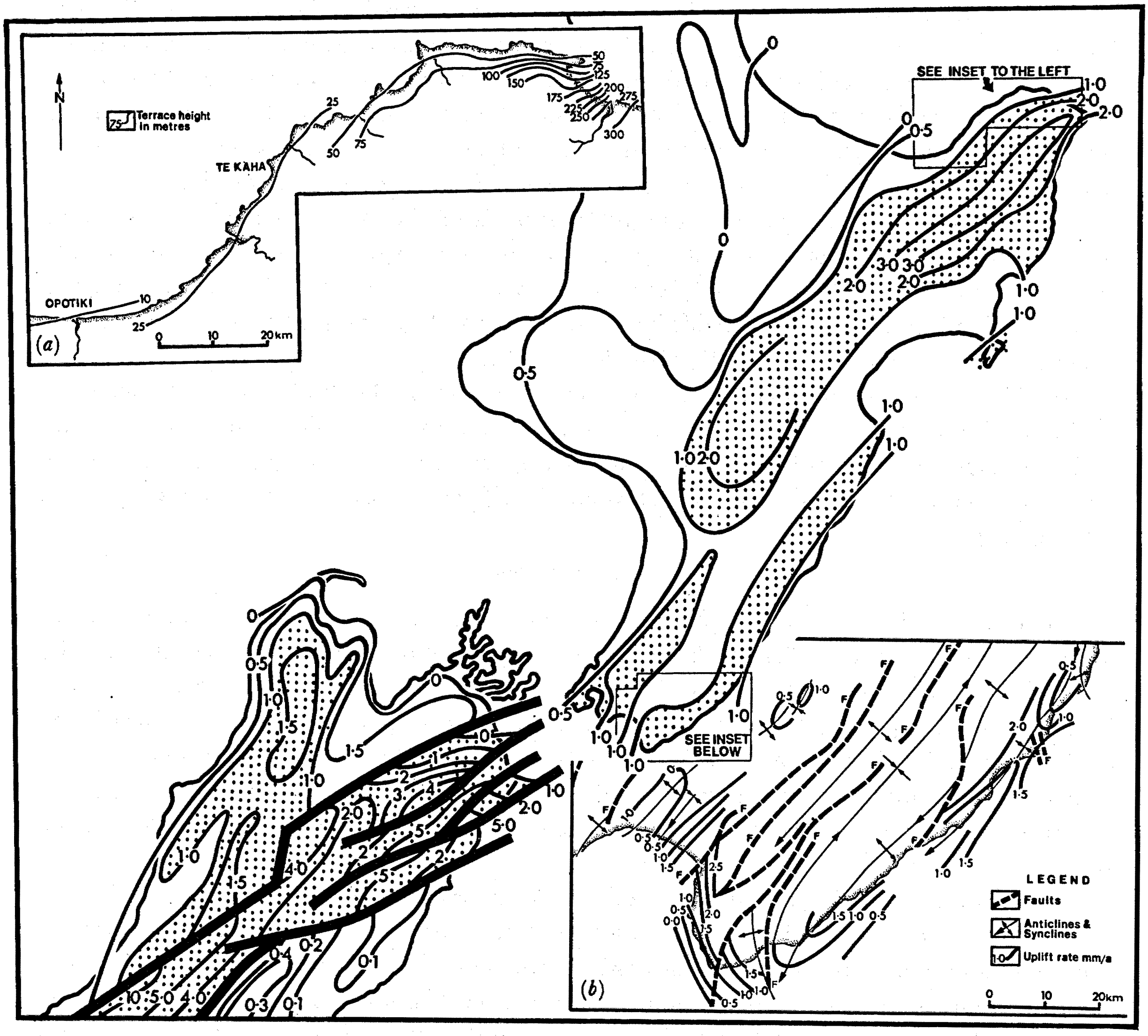

FIGURE 8. Uplift map of the Hikurangi margin derived from Wellman (1979), South Island, and Pillans (1986), North Island. Inset (*a*) top left from Yoshikawa (1980) shows gentle tilting of the 120000 a old marine bench and terrace. Inset (*b*) bottom right from Ghani (1978) shows uplift rates in mm per year from Holocene and last Interglacial bench and terrace heights. Note contrast in style of the major uplifted areas along the axial ranges and along the coastal ranges.

plate (Adams & Ware 1977). A smaller belt of similarly rapid uplift occurs along the east coast of the North Island between Hawke Bay and Cook Strait in the area of the present Coastal Ranges.

There can be little doubt that the cause of the uplift of the Coastal Ranges is crustal thickening through shortening in the fold and thrust belt. Movements of several hundred metres on westerly dipping thrusts and vertical dips in mid-Pleistocene sedimentary strata are observed. Shortening has resulted in marked folding, at a scale of a few kilometres, of the Holocene and last Interglacial marine beaches and terraces (inset to figure 7), (Ghani 1978). The severe internal deformation indicated by the folding and closeset faulting is that expected of an actively deforming frontal wedge.

In contrast, the main belt of late Quaternary uplift is a broad antiformal dome. This can be seen in the uplifted last Interglacial terrace on the northern coastline of East Cape (inset to figure 8). Detailed studies show gentle, uninterrupted tilting over a distance of about 60 km normal to the trend of the belt (Yoshikawa *et al.* 1980, figure 8). In its broad warping of uplifted surfaces and, except where cut by the transcurrent fault zone, little evidence of internal deformation, the uplift of the axial ranges can be described as a regional warping.

While sediments were accumulating in the Wairoa Basin, the axial ranges were rising to the west. These movements continued into late Quaternary time when the whole region began to rise above sea level so that the only part of the Wairoa Basin presently receiving marine sedimentation is in Hawke Bay. Uplift initiated in the axial ranges in Pliocene times has now extended eastwards to include much of the fore-arc basin.

The Plio-Pleistocene subsidence of the fore-arc basin may be the result of crustal thickening to the west under the axial ranges; the added load would cause a regional downward flexure of the subducted oceanic lithosphere that will extend, because of the strength of the lithosphere, over a distance of 200 km or so and which would be reflected in the surface topography of the frontal wedge.

(*c*) *Underplating*

The crustal thickening and uplift of the axial ranges, I suggest, are caused by an accumulation of sediments derived from the Hikurangi Trough and transported beneath the accretionary prism and the fore-arc basin (figure 9). This process of accretion from below has been called underplating (Platt *et al.* 1985). That the transport of sediments beneath the accretionary prism is probable has been shown by the budgetary calculation given earlier. Apart from the uplift, however, there is no direct evidence that underplating is occurring; the excess sediments could conceivably be subducted into the mantle and, therefore, lost to the system.

The rate of underplating required to produce the observed rate of uplift depends on the rate of erosion and the degree of compensation. The deeply dissected nature of the peneplain in the uplifted fore-arc basin indicates an erosion rate about one half the uplift rate. Therefore the rate of change in volume of the uplift (70 km wide, 3 mm a^{-1} peak value) is about one quarter of the rate at which sediments are accreted to the subduction zone (70 mm a^{-1}, 3 km

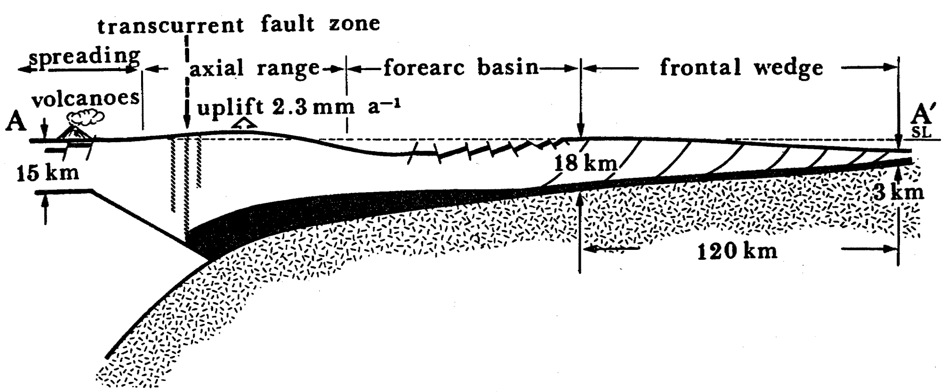

FIGURE 9. Speculative profile of the northern Hikurangi margin showing (in black) an underplated accumulation of sedimentary rocks derived from the trench. The line of section is through the Mahia peninsula (figure 3*a*). Normal faulting in the fore-arc basin has been established only north of the line of section but is presumed to exist throughout the fore-arc basin.

thick). The ratio of the mass of underplated material to mass uplifted could lie in the range 1 (if uncompensated) to 5 (if fully compensated). Therefore to explain the uplift in terms of underplating we require underplating of between one quarter and one and a quarter times the amount of sediments available through subduction. On this basis the hypothesis can be said to be feasible.

In the last 40 Ma about 1000 km^2 per kilometre strike length could have been accumulated and with a width of 100 km suggests an average thickness of 10 km. Uplift of between 2 and 10 km is implied.

5. Discussion and conclusions

The principal evidence for the low-temperature–high-pressure regional metamorphism at the Hikurangi margin is the uplift during the late Cainozoic of the axial ranges and, more recently, the fore-arc basin. Beneath these areas the top of the subducted plate lies at depths of 20–30 km and it is only 2–3 Ma since it commenced its descent at the trough; so, pressures of 6–8 kbar and temperature of 200–400 °C are to be expected. It is the recent uplift of the fore-arc basin that is particularly significant. The northern, older part now lies well above sea level as a raised, deeply dissected peneplain. Yet it is cut and extended by normal faulting some of which is active today. It is difficult to envisage any mechanism other than underplating (the accumulation below of sediments derived from the subducted plate) that would produce uplift while the surface is tectonically stretched and thinned.

In its high heat flow and the substantial volume of intruded igneous rock the Central Volcanic Region is a very likely setting for high-temperature–low-pressure regional metamorphism. However, it is unlikely to be preserved unchanged in the geological record because of its very thin crust. Without compression and restoration of a crustal thickness around 30 km the Central Volcanic Region would be either submerged below sea level on the cooling of the upper mantle or covered by up to 15 km of sedimentary rock. It is conceivable that compression and closure could occur very rapidly. There is presumably a balance between the forces of spreading and the forces involved in the subduction zone, particularly along the decollement above the subducted plate. If the decollement became locked, as it is further south, in the vicinity of Wellington then the overlying crust could be rafted westward at the rate of relative plate convergence, 60 mm a^{-1}, so that in only 1 Ma the Central Volcanic Region could close completely.

The wider implication of the Central Volcanic Region is that some fossil metamorphic belts may, too, have gone through an initial opening phase. This is certainly the most readily understood mechanism by which heat may be rapidly pumped into the upper levels of continental crust. To do so requires the generation of substantial volumes of magma, presumably by partial melting of the depressured asthenosphere, and its injection into the lower continental crust to further produce the more highly silicic melts. The important factors, judging by the Central Volcanic Region, are very substantial spreading in lithosphere that has already elevated isotherms.

As an example of a paired metamorphic belt in the making, the North Island of New Zealand shows several features of importance in inferring the tectonic setting and history of fossil belts.

(i) It is the very rapid change in tectonic style with time that is most impressive. The mountains of southern and eastern North Island are less than 0.5 Ma old (Ghani 1978) and there are several episodes of mountain building and subsidence in the late Cainozoic stratigraphic record that indicate a complicated and poorly understood topographic history. The transcurrent fault zone of the North Island is probably less than 5 Ma old, as it is related to the oblique motion of the plates relative to the plate boundary zone which is a recent development (figure 4). The volcanic rocks of the Central Volcanic Region are all less than 1 Ma old. Although there is a 20 Ma history of volcanism (and probably metamorphism) related to the subduction, it is likely that it is made up of numerous superimposed relatively short-lived events.

(ii) The deformation occurs over a very broad zone several hundred kilometres wide and involves the whole thickness of the crust above the subducted plate. Within this region there are great differences in tectonic style occurring at the same time: back-arc spreading and continental rifting, broad crustal upwarping, transcurrent faulting, fore-arc basin development involving extension and fold and thrust tectonics involving shortening.

(iii) Finally, in its geologically brief existence the Hikurangi margin has been disrupted, separated and a substantial part has been rotated through 90° relative to its hinterland.

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References

Adams, R. D. & Ware, D. E. 1977 Subcrustal earthquakes beneath New Zealand: locations determined with a laterally homogeneous velocity model. *N. Z. Jl Geol. Geophys.* **20**, 59–83.
Bibby, H. M. 1975 Crustal strain from triangulation in Marlborough, New Zealand. *Tectonophysics* **29**, 529–540.
Bibby, H. M. 1981 Geodetically determined strain across the southern end of the Tonga–Kermadec–Hikurangi subduction zone. *Geophys. Jl R. astr. Soc.* **66**, 513–533.
Chase, C. G. 1978 Plate kinematics: the Americas, East Africa and the rest of the World. *Earth planet. Sci. Lett.* **37**, 353–368.
Cole, J. W. 1981 Genesis of lavas of the Taupo Volcanic Zone, North Island, New Zealand. *J. Volcanol. Geotherm. Res.* **10**, 317–337.
Ghani, M. A. 1978 Late Cenozoic vertical crustal movements in the central part of New Zealand. *N. Z. Jl Geol. Geophys.* **21**, 117–125.
Haines, A. J. 1979 Seismic wave velocities in the uppermost mantle beneath New Zealand. *N. Z. J. Geol. Geophys.* **22**, 245–257.
Haines, A. J. 1982 Calculating velocity fields across plate boundaries from observed shear rates. *Geophys. Jl R. astr. Soc.* **68**, 203–209.
Hatherton, T. 1970 Upper mantle inhomogeneities beneath New Zealand: surface manifestations. *J. geophys. Res.* **75**, 269–284.
Hull, A. G. 1985 Late Quaternary geology of the Cape Kidnappers Area, Hawke's Bay, New Zealand. M.Sc. thesis, Victoria University of Wellington.
Karig, D. E. 1970 Kermadec Arc–New Zealand tectonic confluence. *N. Z. Jl Geol. Geophys.* **13**, 21–29.
Katz, H. R. 1982 Plate margin transition from oceanic arc-trench to continental system: the Kermadec–New Zealand example. *Tectonophysics* **87**, 49–64.
Kingma, J. T. 1964 Sheet 9 Gisborne (1st edn). *Geological map of New Zealand* 1:250,000. Wellington, N.Z.: Department of Scientific and Industrial Research.
Lewis, K. B. 1985 *New seismic profiles, cores and dated rocks from the Hikurangi margin, New Zealand.* (*N.Z. Oceanographic Field Report no.* 22) (53 pages.) Wellington, N.Z.: Department of Scientific and Industrial Research.
Mazengarb, C. 1984 The Fernside Fault: an active normal fault, Raukumara Peninsula, New Zealand. *N. Z. Geol. Surv. Rec.* **3**, 98–103.
Minster, J. B. & Jordan, T. H. 1978 Present-day plate motions. *J. geophys. Res.* **83**, 5331–5354.
Mumme, T. M. & Walcott, R. I. 1985 Paleomagnetic studies at Geophysics Division 1980–1983. Department of Scientific and Industrial Research, Geophysics Division Report no. 204. (62 pages.)
O'Halloran, N. G. 1963 Final report of a seismic survey in the Hawke's Bay area of New Zealand. GR24. B.P., Shell and Todd Petroleum. (*New Zealand Geological Survey Report PR*325).

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Pettinga, J. R. 1982 Upper Cenozoic structural history, coastal southern Hawke's Bay, New Zealand. *New Zealand Jl Geol. Geophys.* **25**, 149–191.

Pillans, B. 1986 A Late Quaternary uplift map for North Island, New Zealand. In *Recent crustal movements of the Pacific region. Bull. R. Soc. N. Z.* (In the press.)

Platt, J. P., Leggett, J. K., Young, J., Raza, H. & Alam, S. 1985 Large scale sediment underplating in the Makran accretionary prism, southwest Pakistan. *Geology* **13**, 507–511.

Scholl, D. W., von Huene, R., Vallier, T. L. & Howell, D. G. 1980 Sedimentary masses and concepts about tectonic processes at underthrust ocean margins. *Geology* **8**, 564–568.

Sissons, B. A. 1980 *The horizontal kinematics of the North Island of New Zealand.* Ph.D. thesis, Victoria University of Wellington.

Spörli, K. B. 1978 Mesozoic tectonics, North Island, New Zealand. *Bull. geol. Soc. Am.* **89**, 415–425.

Stern, T. A. 1985 A back-arc basin formed within continental lithosphere: the central volcanic region of New Zealand. *Tectonophysics* **112**, 385–409.

Stock, J. & Molner, P. 1982 Uncertainties in the relative positions of the Australia, Antarctica, Lord Howe and Pacific Plates since the Late Cretaceous. *J. geophys. Res.* **87**, 4679–4714.

Van der Lingen, G. 1982 Development of the North Island subduction system, New Zealand. In *Trench-forearc geology* (ed. J. K. Leggett), pp. 259–272. (576 pages.) *Geological Society of London Special Publication no.* **10**.

Walcott, R. I. 1978*a* Present tectonics and Late Cenozoic evolution of New Zealand. *Geophys. Jl R. astr. Soc.* **52**, 137–164.

Walcott, R. I. 1978*b* Geodetic strains and large earthquakes in the axial tectonic belt of North Island, New Zealand. *J. geophys. Res.* **83**, 4419–4429.

Walcott, R. I. 1984*a* Reconstruction of the New Zealand region for the Neogene. *Palaeogeogr., Palaeoclimatol. Palaeoecol.* **46**, 217–231.

Walcott, R. I. 1984*b* The kinematics of the plate boundary zone through New Zealand: a comparison of short and long-term deformation. *Geophys. Jl R. astr. Soc.* **79**, 613–633.

Walcott, R. I., Christoffel, D. A. & Mumme, T. M. 1981 Bending within the axial tectonic belt of New Zealand in the last 9 Ma from paleomagnetic data. *Earth planet. Sci. Lett.* **52**, 427–434.

Walcott, R. I. & Mumme, T. M. 1982 Paleomagnetic study of the Tertiary sedimentary rocks from the East Coast of the North Island, New Zealand. Department of Scientific and Industrial Research, Geophysics Division Report no. 189. (44 pages.)

Webb, J. H., Wesnousky, S. G. Helmberger, D. V. 1985 A body-wave analysis of the 1966 Gisborne, New Zealand earthquake. *Tectonophysics* **113**, 271–282.

Weissel, J. K., Hayes, D. E. & Herron, E. M. 1976 Plate tectonic synthesis: the relative motion between the Australian, New Zealand and Antarctic continental fragments since the Late Cretaceous. *Mar. Geol.* **25**, 231–277.

Wellman, H. W. 1979 An uplift map for the South Island of New Zealand and a model for uplift of the Southern Alps. *Bull. R. Soc. N. Z.* **18**, 13–20.

Wickham, S. M. & Oxburgh, E. R. 1985 Continental rifts as a setting for regional metamorphism. *Nature, Lond.* **318**, 330–333.

Yoshikawa, T., Ota, Y., Yonekura, N., Okada, A. & Iso, N. 1980 Marine terraces and their tectonic deformation on the Northeast Coast of the North Island, New Zealand. *Geogrl Rev. Japan* **53–4**, 238–262.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES