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The tectonic setting of the Fiordland region, south-west New Zealand

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Summary. The Indian/Pacific plate boundary to the south-west of New Zealand has undergone highly oblique compression during the past 15 Myr. Subduction of the Indian plate north-eastwards under the Pacific plate has probably occurred at Puysegur Trench, south of New Zealand, for most of this period. In contrast, subduction at the plate boundary through south-west New Zealand (Fiordland) has only occurred during the last 7 Myr when oceanic lithosphere of the Indian plate has been adjacent to the continental lithosphere of Fiordland (Pacific plate). Transcurrent motion at the plate boundary during this time has transferred the continental lithosphere forming Challenger Plateau on the Indian plate northwards relative to southern New Zealand. Most geophysical features of the Fiordland region can be accounted for in terms of this recent subduction event at the Fiordland margin. However, the intermediate depth seismicity appears to arise from a fragment of the Indian plate, subducted at Puysegur Trench and subsequently moved northwards by the transcurrent motion between the two plates.

Introduction

The Fiordland region of south-west New Zealand lies along the boundary of the Pacific and Indian plates (Fig. 1) and exhibits elements of convergent and transcurrent plate margins. A problem has always been to relate these elements to a single kinematic model (e.g. Christoffel & Van Der Linden 1972; Hayes & Talwani 1972). Smith (1971) and Smith & Davey (1983) have delineated a steeply dipping planar zone of intermediate depth earthquakes of restricted length underlying this region, indicating subduction of the Indian plate under the Pacific plate to the east. Scholz *et al.* (1973) show a more diffuse zone of intermediate depth hypocentres, possibly because the recording stations used in their survey were fairly closely spaced compared with the depth of the earthquakes. Alternatively, the lower magnitude activity detected by the microearthquake survey may have originated in a wider zone than the larger events. The deepest section of this intermediate depth earthquake zone is displaced to the north-east of the centre of a large dipolar gravity anomaly noted by Woodward (1972). The only calc-alkaline volcanism in the region occurs at Solander Island,

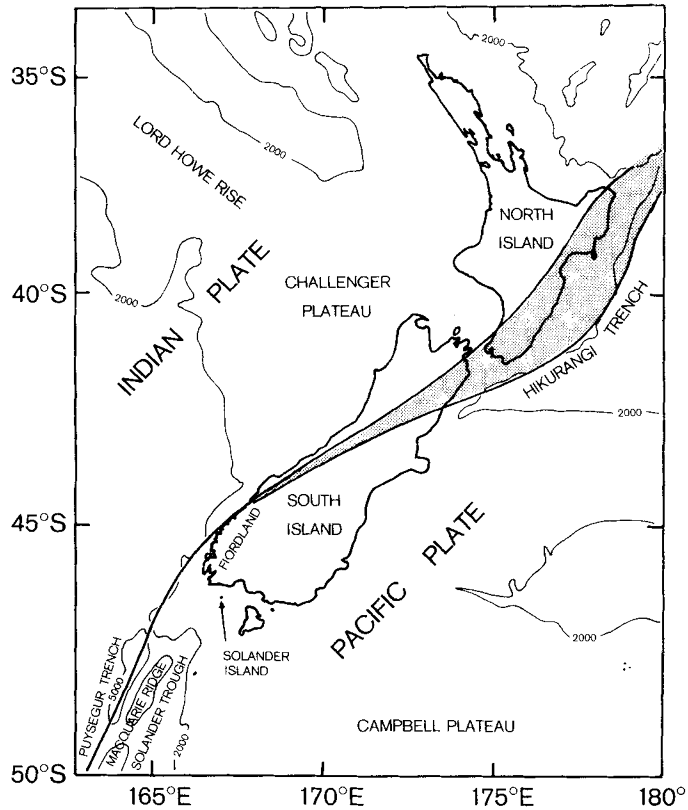


Figure 1. The Indian–Pacific plate boundary (shaded) through New Zealand. Fiordland lies at the south-west end of South Island, New Zealand.

south of Fiordland. The relative motion at the plate boundary in this region during the last 20 Myr has been largely transcurrent with an increasing component of compression (Walcott 1978a). However, no trench is obvious on the ocean side of the boundary at Fiordland (Van der Linden & Hayes 1972) and, to explain the earthquake hypocentre distribution, Christoffel & Van der Linden (1972) have postulated a 'ploughshare' model for the plate boundary with the western (Indian) plate dipping obliquely under the eastern (Pacific) plate at Puysegur Trench with a northwards and clockwise rolling motion.

A significant amount of new geophysical data are now available for the Fiordland region and a reappraisal of the recent tectonic history of this section of the Indian/Pacific plate boundary seems timely. This paper reviews the existing geophysical data and by presenting new data, primarily from offshore, attempts to derive a single dynamic model to fit these data. This is not completely successful and problems with the model are noted.

Relative plate motions

The relative motion between the Indian and Pacific plates in the Fiordland region during the Cenozoic can be computed using the poles and relative rotation rates calculated for these plates. Walcott (1978a) has derived finite poles of rotation for the Pacific and Indian plates back to the time of anomaly 13 (35 Myr BP) using the data of Weissel, Hayes & Herron (1979) and Molnar *et al.* (1975) for the Pacific–Antarctic and Indian–Antarctic plate boundaries. Few of these data were obtained from near New Zealand.

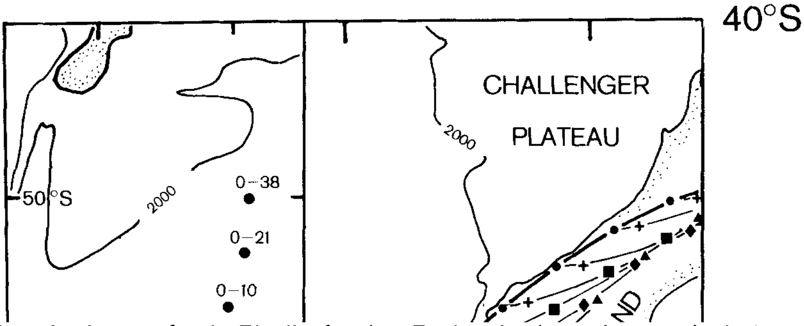


Figure 4. Crustal epicentres for the Fiordland region. Earthquakes larger than magnitude 5 are shown by the larger solid circles. Inset are first motion solutions for large ($M > 5.3$) earthquakes as shown by A–D, solutions E–J are after Scholz *et al.* (1973). Locations marked are BB, Big Bay; MS, Milford Sound; GS, George Sound; DOS, Doubtful Sound; DUS, Dusky Sound.

Crustal velocity structure

Seismic refraction measurements have been made along three profiles over the onshore gravity high by Davey & Broadbent (1980). They show high seismic velocities, up to 7.3 km s^{-1}, at a depth of 8 km. One profile perpendicular to the strike of the gravity high and along Doubtful Sound is shown in Fig. 5. No mantle velocities (> 8.0 km s^{-1}) were recorded but if mantle material has a seismic velocity of 8.4 km s^{-1} (as suggested by Haines 1977 and Smith & Davey 1983) then it must be at least 18 km deep. The high crustal seismic velocity of 7.3 km s^{-1} is at an unusually shallow depth for continental active margins. The depth to the high velocity rocks increases rapidly offshore to the west and more gently inland towards the east (Fig. 5). Seismic refraction data from sonobuoys (Table 1) show that thick sediments, exceeding 4 km in places, occur at the base of the continental slope. These thick sediments coincide with highly folded sediments as shown on seismic reflection data (Fig. 6). The migrated reflection section shows young flat-lying sediments abutting the highly folded and faulted sediments. Further inshore no internal reflections can be detected in the sedimentary sequence, indicating a high degree of tectonic disturbance. Possible low angle thrust-faulting in the folded sequence has been indicated on the section. This record

ment of the Indian plate relative to the Pacific plate since the time indicated by the symbols. The present plate boundary is indicated by solid circles and points at different times corresponding to the present points are linked by flow lines. The positions of the finite poles of rotation are shown by solid circles in the inset map.

The motion of the Pacific plate relative to the Indian plate (Fig. 2) has been calculated for the region using the method of Smith (1981) and the rotation poles of Walcott (1978a). These calculations show a highly oblique compression between the two plates in the Fiordland region during the last 15 Myr, preceded by a small amount of oblique extension. The relative motion in this region is becoming less oblique with time, resulting in the component of convergence normal to the plate boundary becoming progressively larger. This arises from the south-westerly movement of the rotation pole (Fig. 2).

The calculated amount of convergence is greatest at the southern end of the intermediate depth earthquake zone underlying Fiordland where a total of 170 km of Indian plate, measured normal to the present plate boundary at this position, has apparently been consumed. The large amount of concurrent dextral transcurrent movement (Fig. 2), however, suggests that the subducted lithosphere under Fiordland was in fact subducted much further south at Puysegur Trench and has been translated north.

These calculations indicate that a sufficient amount of the Indian plate has been subducted at Puysegur Trench to underlie and be the source of the calc-alkaline andesitic volcanic centre forming Solander Island. If the Solander volcanics were derived from the subducted slab then these relative plate motion calculations indicate that the volcanics must be less than about 4 Myr old. This is consistent with the K-Ar age of 1 Myr recently obtained for these rocks by C. J. D. Adams (private communication).

The results presented in Fig. 2 also show that the continental lithosphere of Fiordland lay adjacent to the continental lithosphere of Challenger Plateau prior to about 7 Myr BP. The relative northwards movement of Challenger Plateau has meant that the plate margin in Fiordland has, from the south, become progressively exposed to the oceanic lithosphere of the Indian plate (Tasman Sea) with time. During the time that oceanic lithosphere has existed on the western side, the convergence across this portion of the plate boundary would vary from about 50 km in the south to zero in the north where Challenger Plateau currently abuts New Zealand. This suggests that the result at Fiordland of convergence across the plate boundary may, in simple terms, be considered in two parts: (1) the oceanic lithosphere of the Indian plate subducted at Puysegur Trench to the south and transferred northwards by transcurrent motion and (2) the recent subduction at the Fiordland coast since it has become adjacent to oceanic Indian plate lithosphere. In fact there will be a continuum between these two cases. Prior to the existence of oceanic lithosphere on the western side of the plate boundary, the convergence between the plates was presumably taken up by uplift and crustal thickening. There are few data available to indicate uplift rates in Fiordland. The data which are available are restricted to marine beaches along the southern coast of Fiordland (Wood 1960). They suggest a low uplift rate of about 450 m during the past 4 Myr (Wellman 1979).

Seismicity

The intermediate depth earthquakes in the Fiordland region can be divided into three groups (Smith & Davey 1983). Most occur in a planar zone approximately 15 km wide dipping at 80° to the south-east with a strike of N40°E and extending from Dusky Sound to George Sound (Fig. 3b). The maximum depth of the main group of hypocentres increases from about 50 km in the south-west to 150 km in the north-east. To the north-east of Milford Sound and to the south-east of Dusky Sound are two isolated groups of intermediate depth hypocentres, the northern group forming a compact cluster of about 20 km diameter at about 60 km depth and the southern group defining a line dipping to the east.

The distribution of the main group of subcrustal earthquakes confirms that at least 150 km of Indian plate has been subducted along this plate boundary. However, the striking decrease in maximum depth of the hypocentres from north to south is too great to be explained in terms of a decrease in the distance to the pole of rotation. Smith & Davey (1983) suggest that the subducted lithosphere slab has broken off from the Indian plate and been rotated into a near vertical position when it impacted the thick continental lithosphere underlying southern New Zealand. They consider the slab to be broken because of the steep dip of the Benioff zone, 80°, which extends to within 50 km of the surface. Other subduction zones dip at up to about 90°, for example in the Marianas Islands region (Isacks & Barazangi 1977), but start at shallow angles and gradually steepen their dip. The shoaling of the intermediate depth activity southwards may reflect the nature of the break in the subducted slab or, alternatively, the lack of stress in this part of the slab after it broke off or was highly folded. Seismicity data south of New Zealand are not good enough to define the present subducted slab under northern Macquarie Ridge and Puysegur Bank. However, the southern group of intermediate depth earthquakes may mark the leading edge of the southern, unbroken, subducted Indian plate. The shoaling may also partially arise from uplift of the southern end of the slab to give the gravity high and outcropping lower crustal rocks seen in this region.

In Fig. 3 we show the areal distribution of intermediate depth epicentres relative to the outcrop geology (Oliver & Coggon 1979) and the extent of the Fiordland gravity high delineated by the $1500\,\mu\mathrm{N\,kg^{-1}}$ Bouguer anomaly contour. The granulitic facies meta-gabbroic diorite (of the Western Fiordland series), postulated by Oliver & Coggon (1979) to be a lower crustal group of rocks, coincide closely with the extent of the intermediate epicentres and their southern limit closely with the gravity high. The southern limit of the main group of Fiordland intermediate depth hypocentres coincides closely with the southern limit of the western Fiordland lower crustal rocks which are separated from the amphibolite facies meta-sediments of south-west Fiordland by the Dusky Fault. A vertical throw on this fault of at least 1 km, down to the south-west was suggested by Davey & Broadbent (1980) from seismic refraction measurements. This coincidence of the postulated lower crustal rocks, the intermediate depth epicentres and the gravity high could suggest that a large degree of uplift of the Western Fiordland block, perhaps as much as 5 km, took place within the last 5 Myr, probably as a result of the breaking off of the initially subducted part of the downgoing slab. If the rotation of the subducted slab to a near vertical attitude occurred on a shorter time-scale than the subduction process the centre of mass of the slab would remain at approximately a constant depth and its upper edge would be raised.

Epicentres for crustal earthquakes (depths less than 33 km) located in the region by the New Zealand seismological network from 1966, when the Milford Sound station was installed, to 1978 inclusive, are shown in Fig. 4. Earthquakes larger than magnitude 5 are indicated and are fairly uniformly distributed amongst the smaller events. The two obvious clusters of events include swarms, or aftershock sequences, associated with the magnitude 7.0 Milford Sound earthquake of 1976 May 4 and with the magnitude 5.9 earthquake of 1974 September 20 located south of Big Bay. Overall, the crustal seismicity in south-west New Zealand defines a zone trending parallel to the coast and extending from about 80 km inland to 80 km offshore. The offshore activity dies away north of Big Bay where the thicker continental crust on the Indian plate is reached. To the south the zone of crustal activity continues over Puysegur Bank and western Solander Trough. The data indicate that the crustal seismicity may increase south of South Island.

Focal mechanisms for crustal earthquakes are also shown in Fig. 4. Some are given by Scholz *et al.* (1973) with additional ones for the Milford 1976 May 4 earthquake and three other events with magnitudes greater than 5.3 for which some teleseismic data are available. The figure shows lower hemisphere plots with solid circles indicating compressional first motions, and open circles indicating dilations. These focal mechanisms may be interpreted as follows. The Milford 1976 May 4 earthquake, A, shows dextral movement at 067° with underthrusting to the south-south-east at 70° dip. The 1968 earthquake at the entrance to Milford Sound B, is consistent with dextral movement at 080° with underthrusting to the SSE at a dip of 67°. The 1972 offshore earthquake, C, is consistent with dextral movement at 105° and underthrusting to the SSW at 80°. The 1974 September 20 Big Bay earthquake, D, is consistent with dextral movement at 076° with overthrusting to the SSE at 70°.

These new focal mechanisms although poorly constrained, indicate a degree of strike-slip motion in the coastal region as well as the underthrusting seen in the results of Scholz *et al.* (1973). Overall, the first motion studies support underthrusting of the Indian plate under coastal Fiordland as far as Haast in the north with a strike of north to north-east. Further inland the data of Scholz *et al.* (1973) show predominantly strike-slip motion.

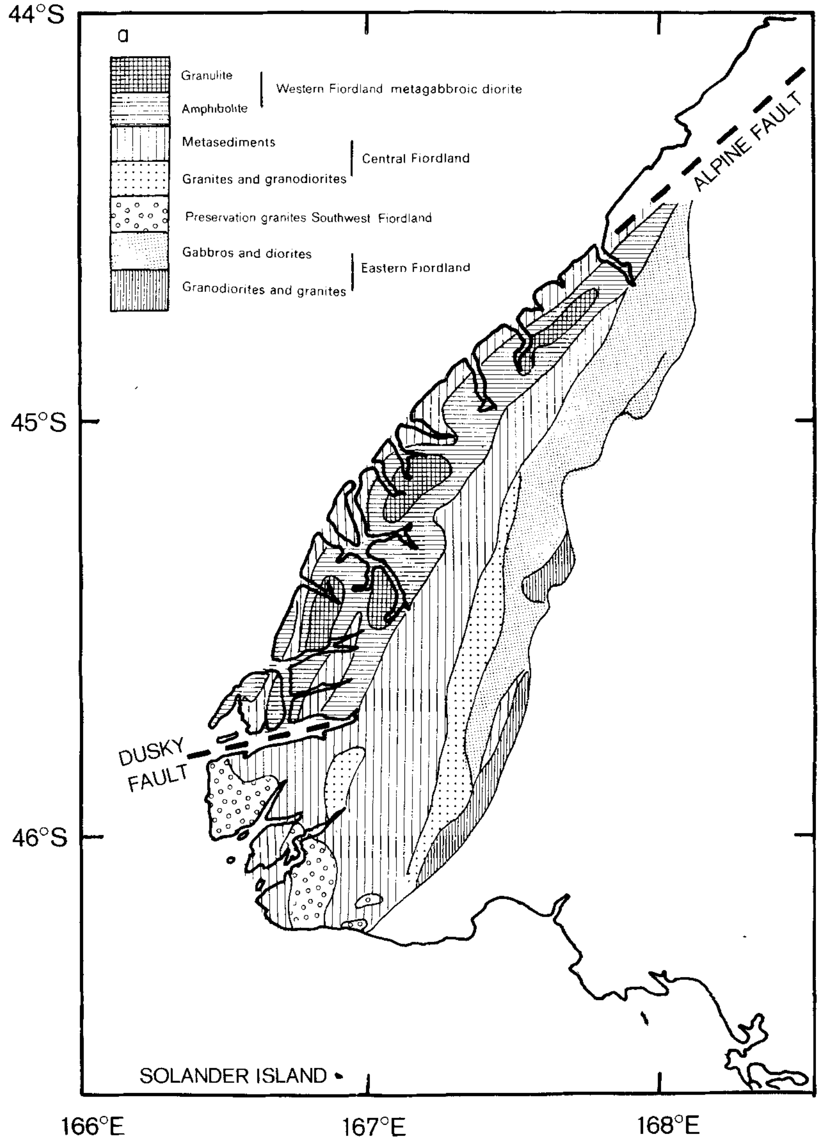


Figure 3. (a) Outcrop geology for the Fiordland block after Oliver & Coggon (1979). The thick dashed lines show the positions of the Alpine Fault and Dusky Fault.

These observations compare closely with those predicted by the model for oblique convergence proposed by Fitch (1972) where a proportion of slip parallel to the plate margin results in transcurrent movement on the continental side of a zone of plate consumption. The present direction of convergence is at 40° to the strike of the subducting plate and would thus, according to Fitch, favour decoupling of the transcurrent slip from movement normal to the plate margin. In Fiordland, however, there are no clearly evident major transcurrent faults on which this continental transcurrent movement is concentrated, although Norris & Carter (1980) suggest limited strike-slip movement along their Moonlight-Hollyford

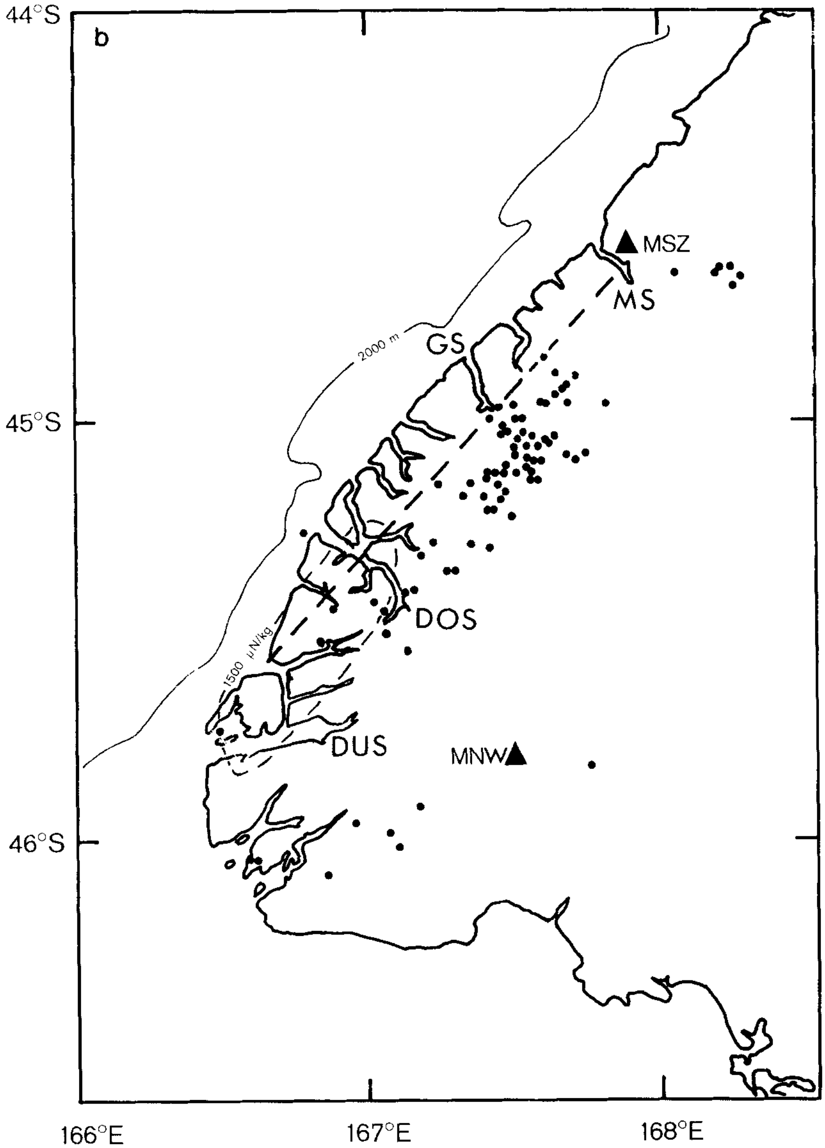
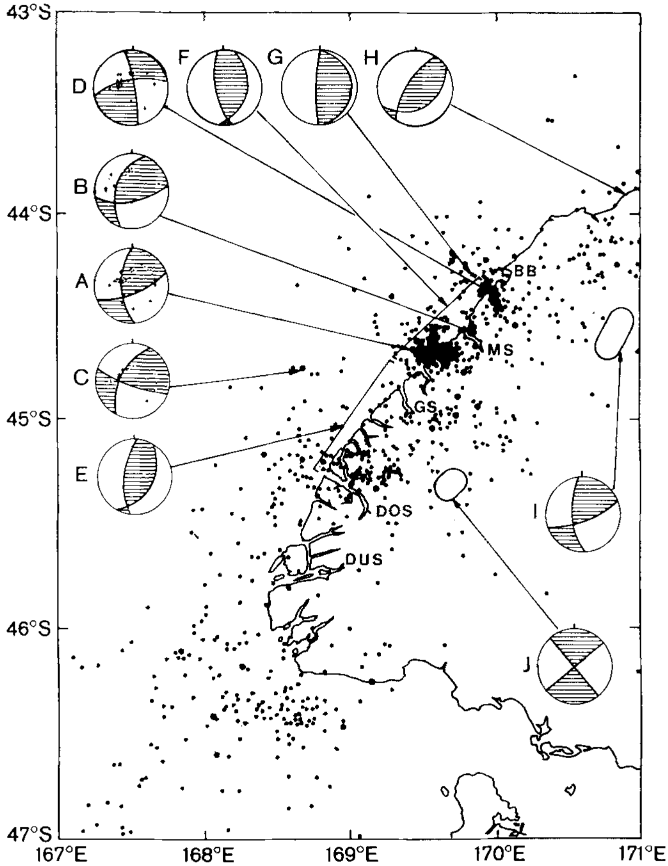


Figure 3. (b) Intermediate depth epicentres are shown by dots and the surface projection of the best fit plane through the main group of hypocentres is shown by the dashed line. The 1500 μN kg^{-1} Bouguer anomaly contour is shown by the fine dashed line. Locations marked are: MS, Milford Sound; GS, George Sound; DOS, Doubtful Sound; DUS, Dusky Sound. Solid triangles MSZ and MNW mark the location of the Milford Sound and Monowai seismograph network stations.

fault system at the eastern margin of the Fiordland block. The focal mechanisms suggest that there is probably some strike-slip motion at the margin itself but the transcurrent motion is perhaps also partially spread over a region of shear through Fiordland, similar to that suggested by Walcott (1979) of the Southern Alps region to the north. This is supported by the findings of Scholz *et al.* (1973) that microearthquake activity extends as far east as Te Anau.



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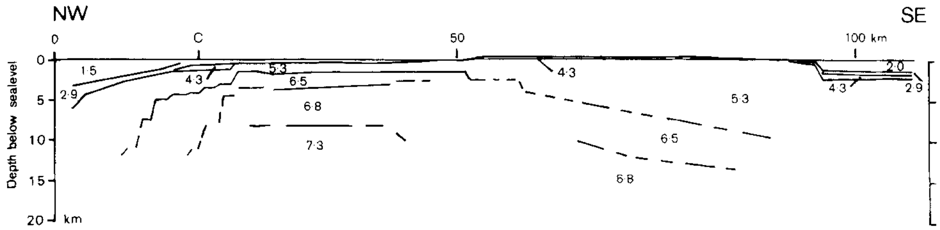

Figure 5. Seismic refraction profile along Doubtful Sound and across the strike of the positive gravity anomaly in Fiordland (after Davey & Broadbent 1980).

shows similarities to seismic data across the young East Luzon and Shikoku island arc systems (Karig & Sharman 1975). The thick sequence of folded and faulted sediments in Fig. 6 would correspond to the accretionary prism and an incipient upper slope basin, the lack of a marked oceanic trench would be due to the young age of the subduction and the thickness of sediments already existing at the base of the continental slope when subduction commenced. The offshore limit of the folded sequence corresponds closely with the western limit of the crustal seismicity. Hatherton (1974) has noted the close correspondence at active margins of one limit of crustal seismicity with the plate boundary at the surface, the trench or ocean edge of the accretionary prism.

Gravity models

The gravity anomalies for the region are shown in Fig. 7. Free air anomalies are shown off-shore and Bouguer anomalies onshore (Reilly & Doone 1972; Woodward, Reilly & Doone 1978). The constraints imposed by the seismicity and seismic refraction results have been used in the construction of models to fit these gravity data. Previous models (Woodward

Table 1. Seismic velocities and layer thicknesses from sonobuoy data.

Buoy	Latitude	Longitude		Layer 1	2	3	4	5	6	7	8
1	45°40′S	165°55′E	v	1.50	1.74*	3.16*	4.97				
			h	4.35	0.38	1.32					
			dip	0.2							
2	45°10′S	166°17′E	v	1.50	(1.75)	2.34	2.67	3.10			
			h	4.15	0.03	0.95	0.44				
3	44°39′S	166°53′E	v	1.50	(1.75)	2.29	2.83	3.63	4.24	4.90	
			h	3.23	0.10	0.82	1.30	2.00	0.42		
			dip		−1.2	−1.5	+1.9	+9	+9		
4	44°16′S	166°57′E	v	1.5	1.78	2.11	2.55	3.18			
			h	3.75	0.31	0.79	0.61				
5	43°58′S	167°36′E	v	1.5	(1.75)	2.27	2.85	3.29	3.99	4.70	6
			h	2.04	0.05	0.50	0.59	1.75	0.34	1.48	
			dip			−1.2					

v, layer velocity in km s^{-1}.
h, layer thickness in km.
dip, dip of layer in degrees, updip is positive, if no value given dip is zero or not available.
( ), assumed velocity.
* interval velocity from variable angle reflection data.

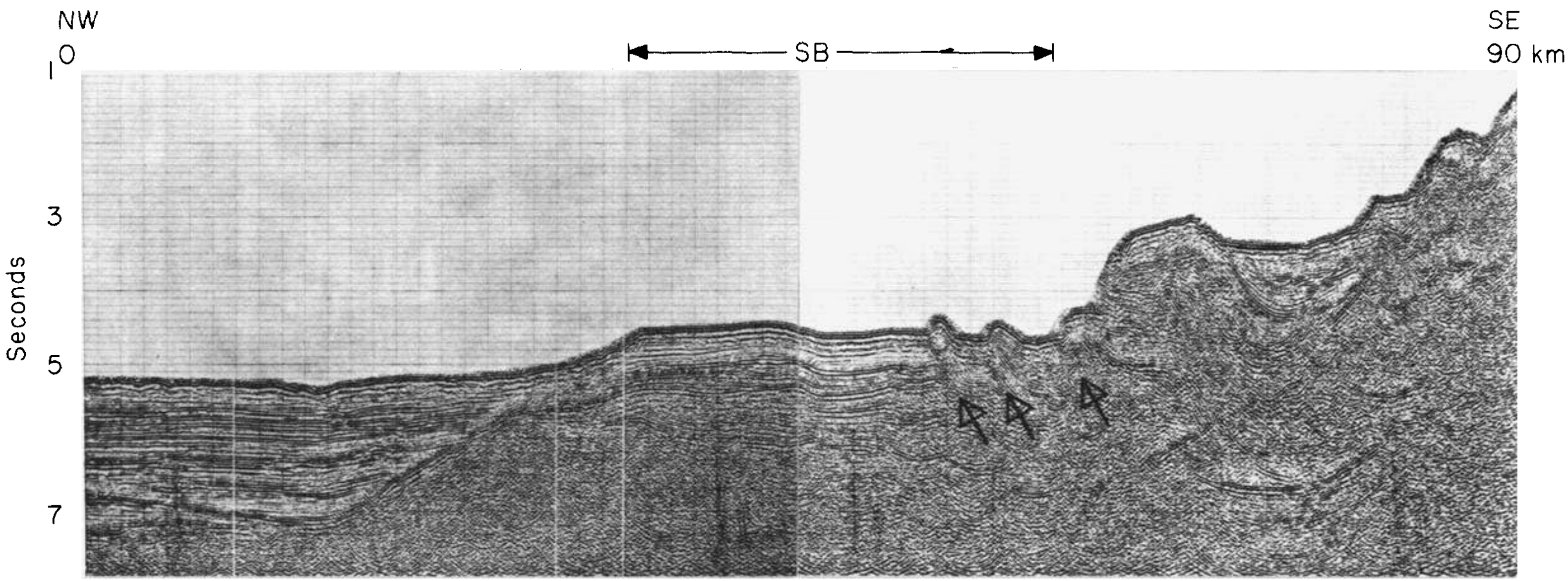

Figure 6. A migrated seismic reflection section across the offshore sedimentary wedge. The profile is located in Fig. 7. Possible thrust faults are indicated. The positions of sonobuoy measurements are shown on the section.

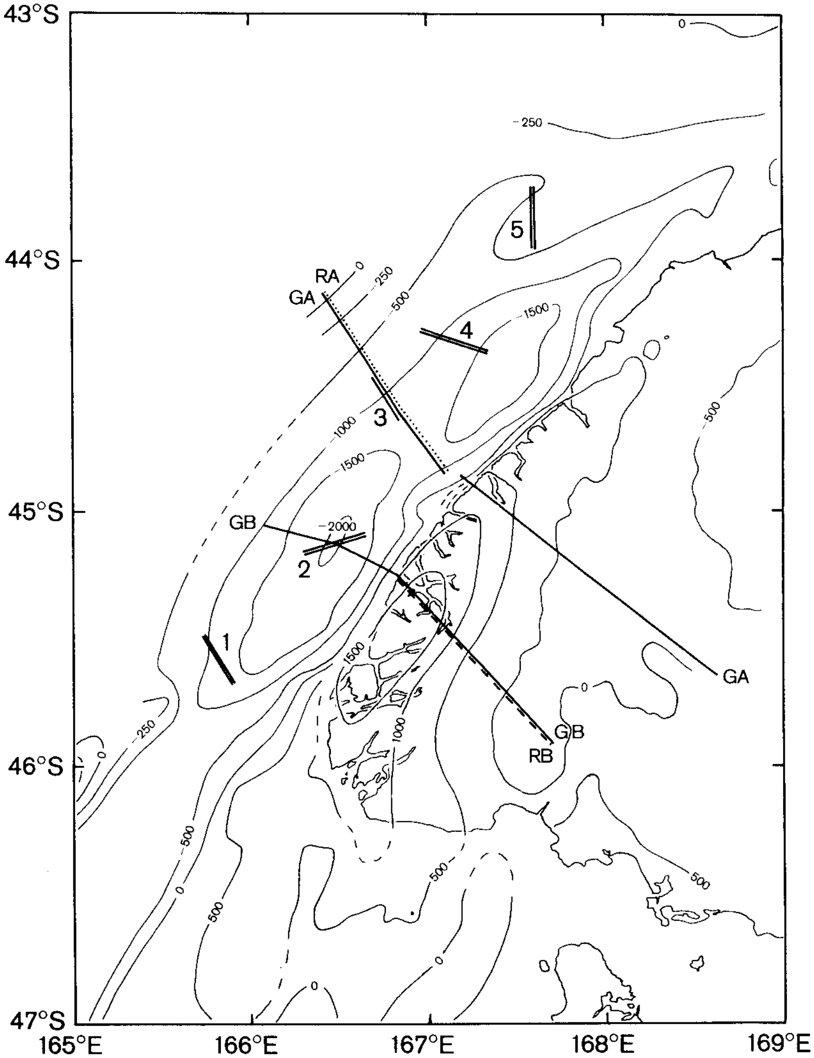

Figure 7. A gravity anomaly map for the Fiordland region. Free air anomalies are shown offshore and Bouguer anomalies onshore. Contour interval is $500\,\mu\mathrm{N\,kg^{-1}}$. The positions of the gravity profiles (Fig. 8) are shown by the solid lines labelled GA and GB, the seismic refraction profile (Fig. 5) is shown by the dashed line labelled RB, the seismic reflection profile (Fig. 6) is shown by the dotted line labelled RA, and the sites of the sonobuoy measurements (Table 1) are indicated by the double lines labelled 1–5.

1972; Walcott 1978b; Oliver & Coggon 1979) lacked all or part of these constraints. Two-dimensional models were constructed for two profiles across the plate margin, one along the seismic profile through Doubtful Sound (Fig. 5), along which the refraction data were more numerous, and one along a profile through George Sound to cross the zone of deepest earthquakes. Both profiles run perpendicular to the strike of the main zone of subcrustal hypocentres. The gravity models are shown in Fig. 8. A density of $2.75\,\mathrm{Mg\,cm^{-3}}$ was chosen for the 5.8 and $6.3\,\mathrm{km\,s^{-1}}$ layers, $3.0\,\mathrm{Mg\,m^{-3}}$ was selected for lower crustal rocks with velocities of 6.8 and $7.3\,\mathrm{km\,s^{-1}}$. Offshore a density of $2.2\,\mathrm{Mg\,m^{-3}}$ was assumed for sediments with velocities of up to $4\,\mathrm{km\,s^{-1}}$ and $2.4\,\mathrm{Mg\,m^{-3}}$ for more compacted sediments. A density of $2.9\,\mathrm{Mg\,m^{-3}}$ was assumed for the oceanic lower crustal layer and density of 3.4 and

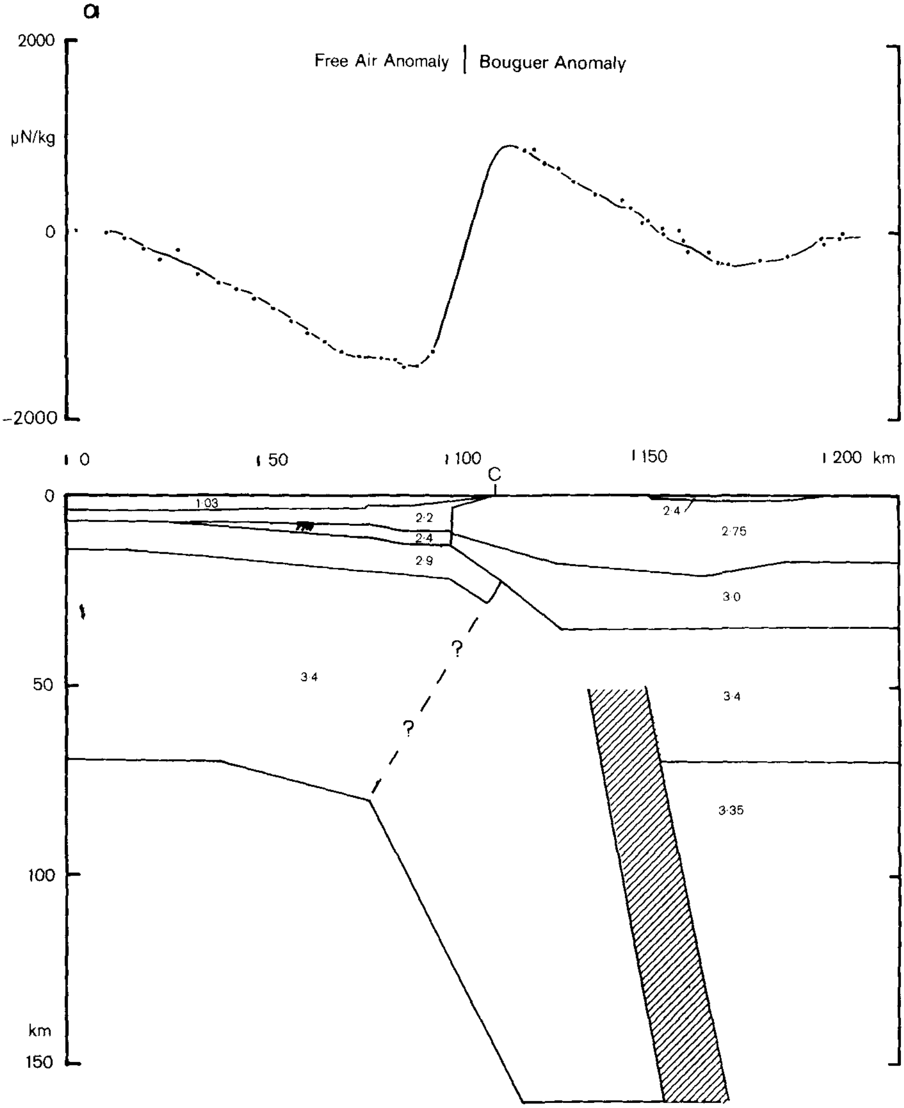

Figure 8. (a) Calculated and observed gravity anomalies and the gravity model for the George Sound profile (GA on Fig. 7). Calculated anomalies – solid line, observed anomalies – dotted line. Densities on the model are in Mg m^{-3}. Seismic boundaries are shown by lines hatchured below. The Benioff zone is shown shaded. C – Coastline.

3.35 Mg m^{-3} for upper mantle (lower lithosphere) and asthenosphere respectively after Grow (1973). The thickness of the oceanic lithosphere, 70 km for an age of 50 Myr (Weissel *et al.* 1979) was derived from the results of Leeds, Karpoff & Kausel (1974). A standard oceanic crustal section 13.5 km thick was assumed for the western parts of the profile and a continental crustal thickness of 35 km assumed to the east. The extent of the low density surface layer to the east was based on surface geology (Wood 1960, 1962, 1966) and the thickness of this layer along the Doubtful Sound profile, profile B, was assumed from the seismic refraction measurements. The oceanic lower crustal layer with a density of

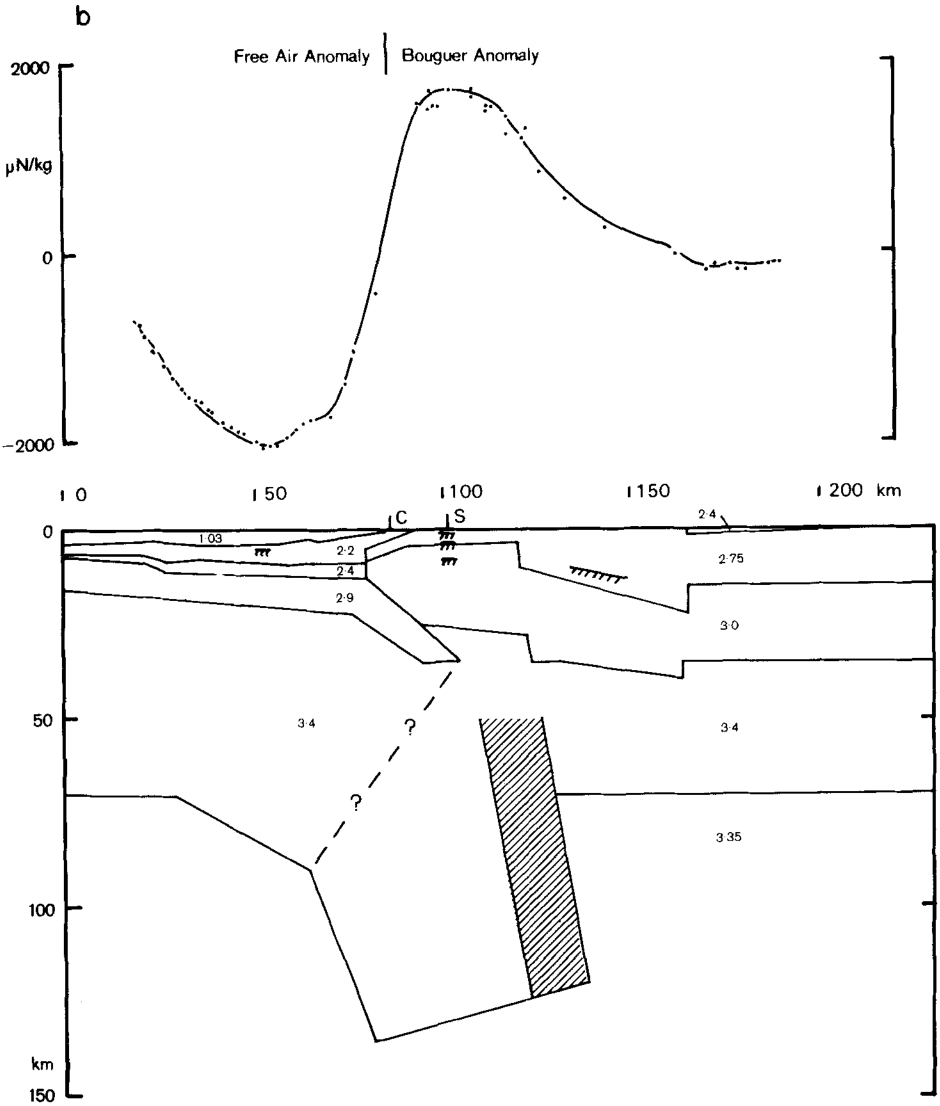

Figure 8. (b) Calculated and observed gravity anomalies and the gravity model for the Doubtful Sound profile (GB on Fig. 7). Other details as for (a).

2.9 Mg m^{-3} has been arbitrarily assumed to have been assimilated into the downgoing lithosphere at a depth of 35 km. This depth is probably a shallowest reasonable value as the basalt-ecoglite transformation occurs within the depth range of 30–60 km.

The oceanic 2.4 Mg m^{-3} layer marks consolidated surface sediments overlying the downgoing slab. These sediments may be in part pelagic sediments scraped from the downgoing oceanic plate but are probably mainly sediments from the southern flank of Challenger Plateau or derived from the high-standing Fiordland block to the east. A mean density contrast of the downgoing slab relative to the surrounding asthenosphere was chosen to be +0.025 Mg m^{-3}. Other workers (Grow 1973; Watts & Talwani 1974) have used density contrasts of about 0.04–0.05 Mg m^{-3} for underthrusting rates of about 80 mm yr^{-1}. In the Fiordland–Puysegur region the underthrusting rate has been about 10 mm yr^{-1} over the last

10 Myr and the density contrast would be expected to be less, perhaps as low as 0.01 Mg m^{-3} (Oxburgh & Turcotte 1970; Minear & Toksöz 1970). A value of 0.025 Mg m^{-3} was selected as a maximum reasonable value.

At about 140 km on the Doubtful Sound profile it was not possible to match the 2.75–3.00 Mg m^{-3} interface on the gravity model with the postulated seismic interface of Davey & Broadbent (1980). However, they note that the interface shown represents a minimum depth for the interface, it could lie at a greater depth and hence be compatible with the gravity model. A simple gravity model was also used so as to minimize the number of density layers; too large a density could have been used in the region of this depth discrepancy especially in view of the Palaeozoic granites and granodiorites which occur in this region (see Fig. 3).

A mjor problem that arises with the gravity model is the relation of the slab defined by intermediate depth earthquakes, hypothesized by Smith & Davey (1983) as having been broken off, to the present subducting slab. The relative plate motion calculations indicate that a maximum of 50 km of subduction of oceanic lithosphere could have occurred across the Fiordland margin on the Doubtful Sound profile and about 25 km of subduction on the George Sound profile. The exact nature of the downwarping of the downgoing lithosphere at the onset of subduction is obviously critically important. Not having any strong evidence to work on, subduction of 40 km for the Doubtful Sound profile and 20 km for the George Sound profile was assumed. The older broken slab is assumed to continue up to 50 km below the surface, as indicated by the intermediate depth seismicity. The junction of the upper part of the broken slab with the lower part of the currently subducting lithosphere is unknown so a simple model has been used. Small changes to this boundary will not affect the gravity models significantly.

The models for both profiles require the western margin of the Fiordland block to be uplifted. In the Doubtful Sound profile, downfaulting of the Fiordland block at its eastern margin is required to fit the gravity anomaly. This eastern edge of the tilted Fiordland block runs along the western margin of the Waiau syncline and through western Lake Te Anau to the Alpine Fault in the north at Big Bay. It presumably coincides with a major fault zone, the Fiordland Boundary Fault (Norris & Carter 1980) which probably joins up with the Hollyford fault system in the north and along which the major lakes of eastern Fiordland lie.

Conclusions

The poles of rotation for the Pacific–Indian plate boundary show that convergence and hence subduction has probably occurred at Puysegur Trench for at least the last 10 Myr. This was highly oblique during the earlier part of this time, becoming progressively more compressive with time. The relative plate motion at the boundary in Fiordland further north has also been initially transcurrent and subsequently more convergent, although the onset of subduction along the Fiordland margin probably migrated progressively northwards relative to the Pacific plate during the past 7 Myr.

The plate motion calculations suggest that the Benioff zone under Fiordland, as defined by the intermediate depth seismicity, has originated from oceanic crust of the Indian plate subducted at Puysegur Trench and subsequently translated northwards as a result of the large amount of transcurrent movement at the boundary. The total covergence normal to the plate boundary over the last 15 Myr would total some 170 km, sufficient to account for the subducted slab under Fiordland as delineated by the intermediate depth seismicity. The southern group of intermediate depth hypocentres may mark the leading edge of the unbroken Indian plate subducted at Puysegur Trench.

The plate motion data also indicate that the Fiordland margin has been exposed progressively to oceanic Indian plate over the last 7 Myr by transcurrent movements on the Indian–Pacific plate boundary. Prior to that time, a continental block, Challenger Plateau, abutted against Fiordland. The development of a subducting margin along the Fiordland coast has thus only been possible over this period and during that time convergence has been highly oblique. Possible subduction along the Fiordland margin would thus have to be young and developing in the thick sediments forming the southern flank of Challenger Plateau supplemented by high sedimentation from the uplifted Fiordland block. Analogies can be drawn with the young subducting margin of East Luzon discussed by Karig & Sharman (1975). The lack of a trench in that region is accounted for by the thick sediments already existing at the plate boundary. The accretionary wedge is quite easy to recognize on the seismic section (Fig. 6) with the upper slope basin being infilled by sedimentation from the uplifted Fiordland block. These sediments have been deformed by the thrust and transcurrent faulting.

Solander Island is the only known andesitic volcanic centre associated with the Fiordland–Puysegur compressive plate margin (Harrington & Wood 1958). The age of the volcanic centre is about 1 Myr. It is located where the amount of convergence is greatest. Fig. 2 shows that convergence between the Indian and Pacific plates decreased to the north and south of the Solander Island section and this perhaps accounts for the lack of andesitic volcanism elsewhere along this sector of the plate boundary.

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