THE APPLICATION OF DEEP SEISMIC REFLECTION PROFILING TO PETROLEUM EXPLORATION: AN EXAMPLE FROM THE SOUTH TARANAKI BASIN

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A deep seismic reflection profile shot across the south Taranaki Basin indicates up to 10 km of crustal thickening beneath the eastern margin of the basin. The seismic data also document a broad (150-200 km half-wavelength) flexure of the whole crust with the locus of the loading, that gives rise to this flexure, appearing to be in the vicinity of the Taranaki Boundary Fault. Such a crustal thickening and flexure is suggestive of a compressional origin for the Oligocene-Miocene subsidence within the basin, rather than an extensional one as previously assumed. A zone of stacked basement thrusts is interpreted to exist for up to 50 km east of the Taranaki Boundary Fault. This zone of stacked thrusts is considered to be the Taranaki Fault Zone. Comparison of deep seismic sections from the South Taranaki Basin and a line that crossed the Pyrenees suggests up to 100 km of shortening may have occurred within the Taranaki Fault Zone. Some of the shortening expressed in the deep crustal deformation may, however, be a relic from a Paleozoic or Mesozoic event. The configuration of overthrust belt, crustal thickening, and a broad crustal flexure filled by flysch-type sediments and older rift-margin sediments, fits the foreland basin concept as propounded by Coney, Beaumont and others. A recognition that the main subsidence phase of the South Taranaki Basin was driven by compression and overthrusting, rather than extension, has important implications for estimating the thermal history and overall prospectivity of the basin.

INTRODUCTION

Deep seismic reflection is the application of standard oil industry seismic methods to the study of the whole of the earth's crust. In continental areas the crustal thickness generally varies from about 20 to 50 km. Therefore the principal difference from oil company seismic surveys is that larger energy sources are used, if possible, and seismographs listen longer for reflected energy, i.e. typically 15-20 s two way travel-time (TWT) rather than 5-6 s.

The main goal of deep seismic has been to further our understanding of the makeup and evolution of the continental crust. The questions addressed have largely been global and at times academic. Recently there has been a more pragmatic move to applying crustal seismic studies to specific questions associated with the deep crustal structure beneath oil-producing sedimentary basins. Because deep seismic lines are often of the order of 100-200 km long, and are generally sampling to 2-3 times the depth that standard industry seismic lines penetrate, a more regional view is gained, which in turn often provides a new perspective on the geohistory of the basin in question.

This paper is directed to a brief description and interpretation of a deep seismic reflection profile from the south Taranaki Basin. In particular, one of the main intents of the paper is to demonstrate how deep seismic data has led to an interpretation of the south Taranaki Basin in terms of a compressionlly-driven, foreland basin structure.

EXTENSIONAL BASINS AND FORELAND BASINS: CRUSTAL STRUCTURE SIGNATURES

Early seismic refraction studies showed the earth's crust to consist of 2 or 3 plane layers of velocity 5 -7 km/s overlying an upper mantle of velocity around 8 km/s. The boundary between the crust and the upper mantle is referred to as the *Moho* after the Yugoslavian seismologist Mohorovicic who discovered it (see Jarchow and Thompson (1989) for a discussion on the evolution of ideas about the Moho). More recent deep seismic reflection studies show that the crust is much more complex than the early seismic refraction results suggest. In particular, the Moho is often marked by a broad transition zone, up to 10 km thick, of high reflectivity and the crust itself displays lateral heterogeneity over distances of only 10-20 km (e.g. Schilt et al., 1979). In reflection seismology the *reflection Moho* is taken as the depth where reflectivity at the base of the crust ceases (Klemperer et al., 1986), as this generally is found to coincide with other geophysical determinations of depth to Moho.

Because the Moho represents a first-order change in seismic velocity, it is also inferred to represent a similarly large increase in density. Therefore, if the Moho is arched upwards, say by stretching the crust; isostasy will require that the earth's surface bend downwards to compensate. This process results in an extensionally-driven basin (McKenzie,1978) and is one of the two principal processes by which sedimentary basins are formed. If, on the other hand, thrust sheets are loaded onto a portion of the lithosphere, the earth's crust will respond isostatically by flexing downward ahead of the thrusts. This is the other end-member sedimentary basin, variously labelled exogeosyncline (Kay, 1951), foredeep or foreland basin (Dickinson,1974; Beaumont,1981); the term foreland basin will be used here.

Fig. 1 shows a cartoon of the crustal structure for the two end member sedimentary basins discussed above. Also shown are cartoons depicting the deep seismic reflection response to these basins. For basins formed by crustal extension, or stretching, the Moho appears flat on the time section because, as Warner (1987) argues, lithospheric strength is low, local isostasy prevails, and therefore both the density and traveltime integral down to the Moho remain constant. In contrast, a foreland basin shows a dipping Moho with the thickest section of the crust immediately adjacent to the overthrust belt; here regional isostasy is distributing the load of the thrust sheets. Hence the simple analysis presented in Fig. 1 suggests that if deep seismic reflection data can track the Moho beneath a sedimentary basin, then it may provide an insight to the driving mechanism for the basin in question.

In this study we apply this simple principle to the South Taranaki Basin. Although the prevalent thinking is that the South Taranaki Basin is a rift structure (e.g. Kamp,1986 and others), the deep seismic data from this area shows at least 10 km of crustal thickening beneath, and a broad crustal flexure down towards, its eastern margin. A compressional origin for the South Taranaki Basin is thus proposed.

STRUCTURAL BACKGROUND

A location map for both the Taranaki Basin and the seismic line discussed here are shown in Fig. 2. Over 25 000 km of oil industry shallow seismic (5-6 s two-way travel-time (TWT)) and more than 20 drillholes have provided the principal geological and structural information about the basin (Pilaar and Wakefield, 1978; Knox,1982; King and Robinson,1988). Geophysical data show that the North Taranaki Basin is different from the South Taranaki Basin structurally (Knox,1982), thermally (Pandey,1981) and in the distribution of volcanics (Hatherton *et al.*,1979). The results from this study, therefore, pertain to the South Taranaki Basin only.

Boundaries for the South Taranaki Basin are not clear cut. For this study the eastern boundary is taken at the Taranaki Boundary Fault as shown in Fig. 1. The Taranaki Fault Zone on the other hand has been regarded as a relatively narrow zone taking in the Taranaki Boundary Fault and fault offshoots like the Manaia Fault (Pilaar and Wakefield,1978). In this study Taranaki Fault Zone refers to a zone of inferred stacked thrusts that takes in the Manaia and Taranaki Boundary Fault and also extends well to the east of these two faults (Fig. 2), as will be discussed shortly. The southern boundary of the South Taranaki Basin is taken as the Northwest Nelson coastline, the northern boundary as the Taranaki coastline (Fig. 2). These are arbitrary choices for the northern and southern boundaries, but will suffice for the present study. To the west the South Taranaki Basin merges into the Western Platform: an area of relatively undisturbed conti-

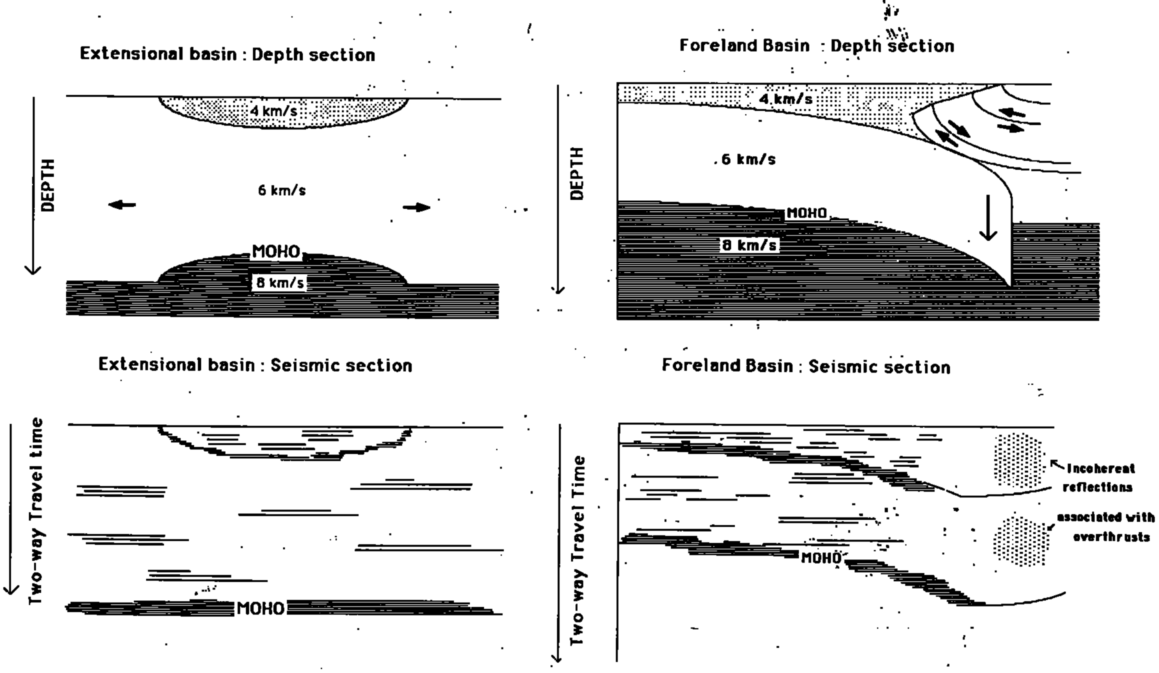


Fig. 1a: Cartoons showing simple crustal structure models for a foreland basin and a basin driven by uniform stretching. Fig. 1b: Cartoons showing the generalised seismic reflection response to the two end-member basins of Fig 1a. Note the flat Moho associated with the extensional basins is due to local isostatic compensation of the sediments by a thinned crust. For the foreland basin the Moho is flexed downwards and compensation for the sediments is regional rather than local.

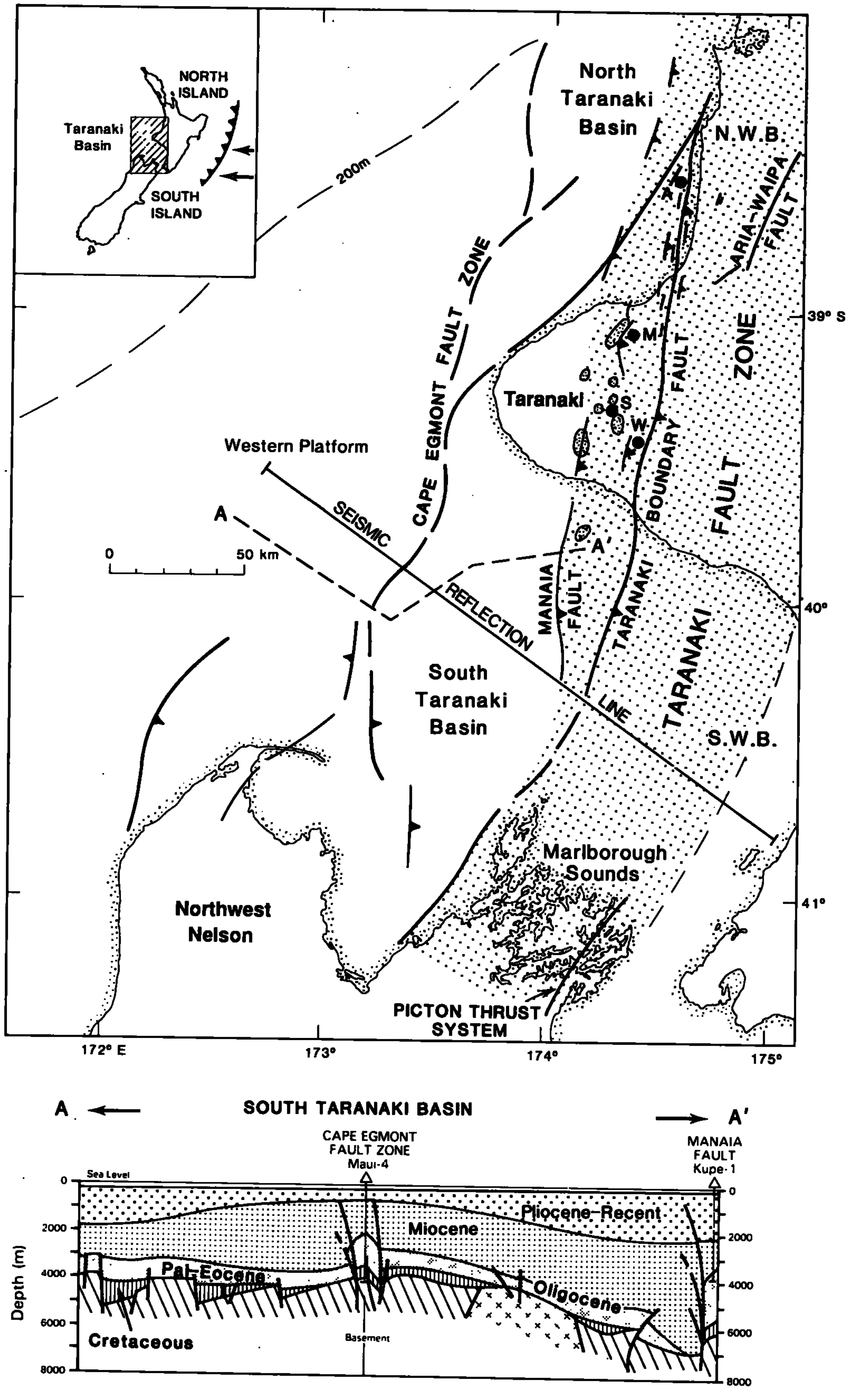


Fig. 2: Location map for the Taranaki Basin showing the location of the deep seismic line, and the drill holes used in the subsidence analysis presented in Fig. 7a. Cross section for the South Taranaki Basin at the bottom is after Pilaar & Wakefield (1978). Dotted areas are discovered oil fields and solid dots are drill holes used in the analysis of Fig. 7a : A=Awakino-1, M=McKee-1, S=Stratford-1, W=Waihapa-1. The coarse dotted area is the estimated extent of the Taranaki Fault Zone as described in the text. NWB=North Wanganui Basin, SWB =South Wanganui Basin. Picton Thrust System after Nicol (1988).

nental shelf dominated by a large thickness of westward prograding Plio-Pleistocene sediments called the Giant Foreset Beds (Pilaar and Wakefield,1978).

A cross-section of the South Taranaki Basin is shown at the bottom of Fig. 2 and demonstrates its asymmetry. Implicit in this cross-section are two distinct phases to the geological history of the Taranaki Basin. From late Cretaceous to Oligocene was a rifting phase linked to the opening of the Tasman Sea (Weissel *et al.*, 1977). From late-Oligocene to late-Miocene a wedge of sediments developed in the Taranaki Basin with a maximum thickness of 3-4 km found adjacent to the Taranaki Boundary Fault (Fig. 1 and Thrasher and Cahill, 1990). It is this wedge-shaped, predominately Miocene package that is proposed to be the foreland basin sequence.

The seismic line crossed the southern portion of the Wanganui Basin and then the South Taranaki Basin (Fig. 2). Anderton (1981) has analysed industry seismic from the Wanganui Basin and describes the basin as a quasi-circular, 4-5 km deep Plio-Pleistocene basin filled with shallow marine sediments. Therefore the Wanganui Basin is a younger feature than the adjacent South Taranaki Basin. A local depocentre containing up to 2 km of Plio-Pleistocene sediments is present within the eastern portion of the South Taranaki Basin (Anderton, 1981; Thrasher, 1990). These may be attributed to distributed flexure linked to Plio-Pleistocene crustal down-warping centred within the Wanganui Basin (Stern and Davey,1989) or they may be an early Pliocene depocenter linked to a late stage of thrusting on the Taranaki Boundary Fault. Therefore in a general sense both Taranaki and Wanganui Basin are related to the development of a subduction zone beneath the North Island, and at present the Taranaki Basin lies behind the Australian-Pacific plate boundary with the subducted Pacific plate being at depths of up to 300 km beneath Taranaki (Adams and Ware, 1977).

DEEP SEISMIC DATA

The deep seismic line was 220 km long, crossed the south Wanganui and south Taranaki Basins (Fig. 2) and was shot by Western Geophysical with the processing carried out by GECO(NZ). Initial interpretations of these data have been presented by Davey (1987) and Stern and Davey (1989).

Figs. 3a and 3b show sections of the seismic data from each side of the Cape Egmont Fault Zone. To the west of the Cape Egmont Fault Zone (Fig. 3a) the crustal section resembles a classic deep crustal section from an area that has undergone extension (e.g. Cheadle *et al.*,1987; Allmendinger *et al.*,1987). That is, a strongly reflective sedimentary section down to about 2.5 s TWT (two way travel-time) and an almost equally reflective crustal section down to 10 s TWT (approximately 30 km deep). At about 10 s TWT the reflectivity ceases and this has been noted world-wide to correspond to the base of the crust, or Moho, where the Moho has been detected by other geophysical means (e.g. Klemperer *et al.*, 1986). East of the Cape Egmont Fault Zone the crustal reflectivity is not as strong and the Moho is marked by a distinct band of reflectors that form a ramp dipping to the east. The apparent crustal thickening to the east is not merely a velocity pull-down effect associated with the thickening wedge of sediments within the upper crust, as the thinnest crust appears to the west of the Cape Egmont Fault Zone where the time delay due to sediments will be the greatest.

Fig. 4 shows a line drawing presentation of the full 220 km long seismic line. Salient points to note on this line drawing are:

(a) The long-wavelength apparent flexure of the crust with the maximum deflection occurring adjacent to the Taranaki Boundary Fault.

(b) Crustal thickening adjacent to the Taranaki Boundary Fault of about 3 s TWT which corresponds to about 10 km in depth.

(c) Beneath the centre of the Wanganui Basin there is a wide *bow-tie* of criss-crossing reflection events at about 10-14 s TWT. Previous work (Davey,1987; Stern and Davey,1989) shows this bow-tie could be resolved by line-segment migration into the westward dipping reflections from subducted Pacific Plate and eastward dipping Moho reflections associated with the Australian Plate.

(d) A zone about 50km wide between the centre of the Wanganui Basin and the Taranaki Boundary Fault where few, if any, coherent deep reflections could be obtained. This is the area that will be referred to here as the Taranaki Fault Zone. In this zone it is proposed the crust is broken by thrust faults that dip steeply eastward into the crust as will be discussed shortly.

STRUCTURAL INTERPRETATION

The Taranaki Fault Zone

Figs. 5a and 5b summarise an interpretation of the deep seismic in terms of a classic foreland basin structure as, for example, outlined by Coney (1973) and Beaumont(1981). It is proposed that the long-wavelength bending of the lithosphere is due to lithospheric flexure induced by the loading of thrust sheets at and east of the Taranaki Boundary Fault. Between the Taranaki Boundary Fault and the Wanganui Basin there are few coherent reflections apart from some strong amplitude reflections, about 40 km from the eastern end of the seismic line, with apparent easterly dips of about 40°. Similar steeply dipping and strong reflections are associated with the Wind River thrust, western USA (Smithson *et al.*, 1981), and the overthrust Arunta block within central Australia (Goleby *et al.*,1989). These reflections have been interpreted to result from shear or mylonite zones associated with *thick-skinned* crustal thrusting. Thus on the interpretation of Fig. 5a a zone of crustal thrusts is interpreted as extending from the middle of the Wanganui Basin westward to the eastern margin of the South Taranaki Basin. It is not clear why few crustal reflections are evident just east of the Taranaki Boundary Fault, but if the thrust zones become greater than about 45° in dip, reflections from them become increasingly difficult to image (Lynn and Deregowski,1981).

Foreland basin model

As shown in Fig. 5b the principal geological characteristic of a foreland basin is that at its base is a sequence of rift margin, or miogeosynclinal, sediments, generally the source rocks for the hydrocarbons, that have been buried by younger flysch-type sediments. Many of the major oil-bearing basins of the world are foreland basin structures. Examples include the Persian Gulf, the Alberta Basin of western Canada, the Ouachita Basin of the southern United States, and the Appalachian Basin of the eastern United States (Oliver,1986; Dickinson,1974).

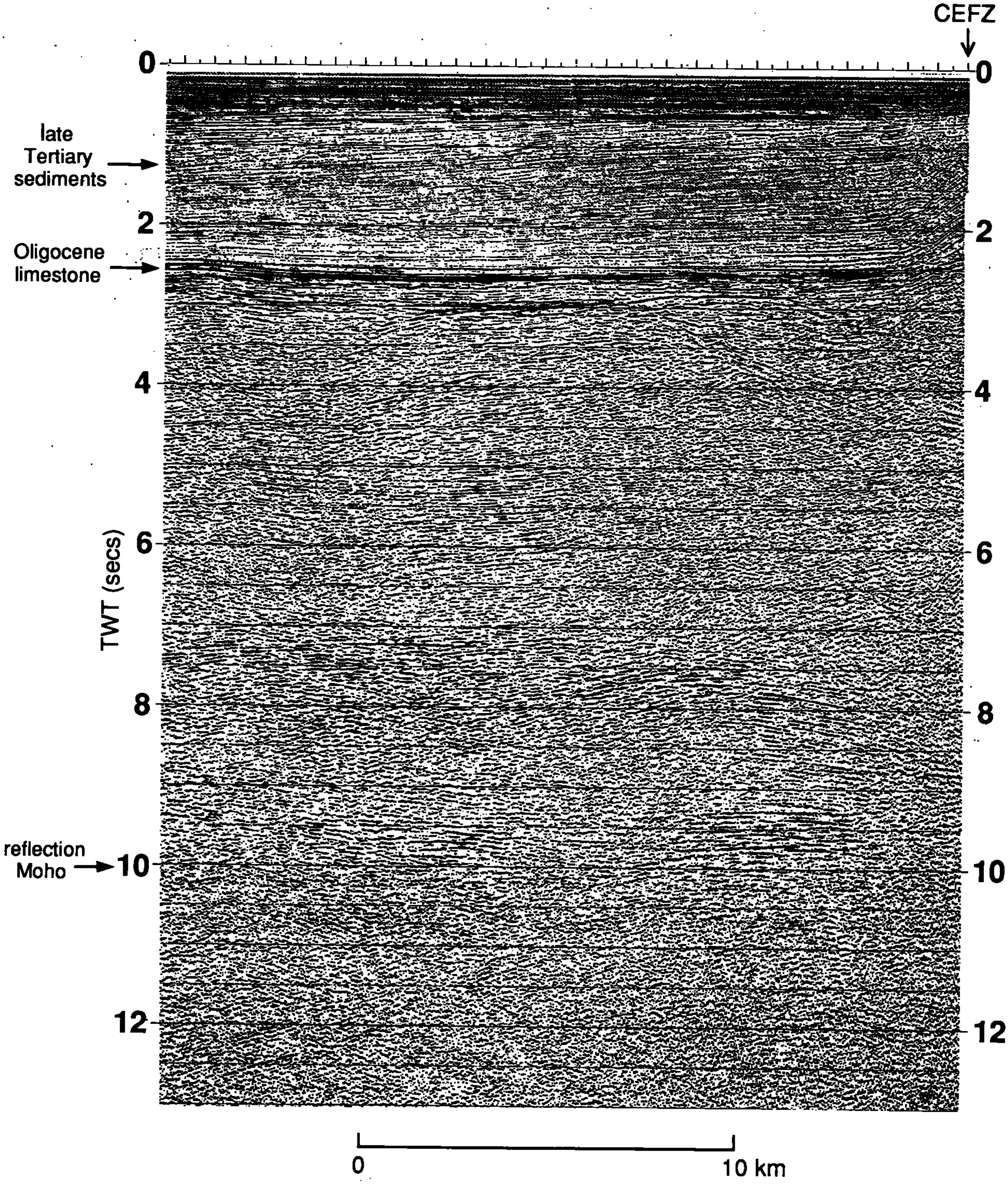


Fig. 3a: Part of the deep seismic data from west of the Cape Egmont Fault Zone. Note the late Tertiary sedimentary section with the westward dipping beds and the prominent Oligocene limestone reflector at 2.5 s TWT. Beneath this reflector there is a strongly reflective crust down to about 10 s TWT where the reflection Moho is picked (roughly 30 km deep i.e. depth(km) ≈ TWT(sec) x 3).

In South Taranaki Basin the Kapuni and Pakawau Formations comprise the rift margin sequence, and the flysch equivalent is represented by the Mahoenui Formation (Pilaar and Wakefield, 1978). However, an interesting departure of South Taranaki Basin from the generic foreland basin model shown in Fig. 5b, and from the above mentioned examples, is that the polarity of the miogeosyncline is reversed, i.e. the late Cretaceous rift-margin on the western side of New Zealand was open to the west, rather than to the east as suggested by the model. In this regard an analogy to the South Taranaki Basin is the Alaskan North Slope where the miogeocline thickens to the north but present-day thrusting is directed from the south (Carman and Hardwick,1983).

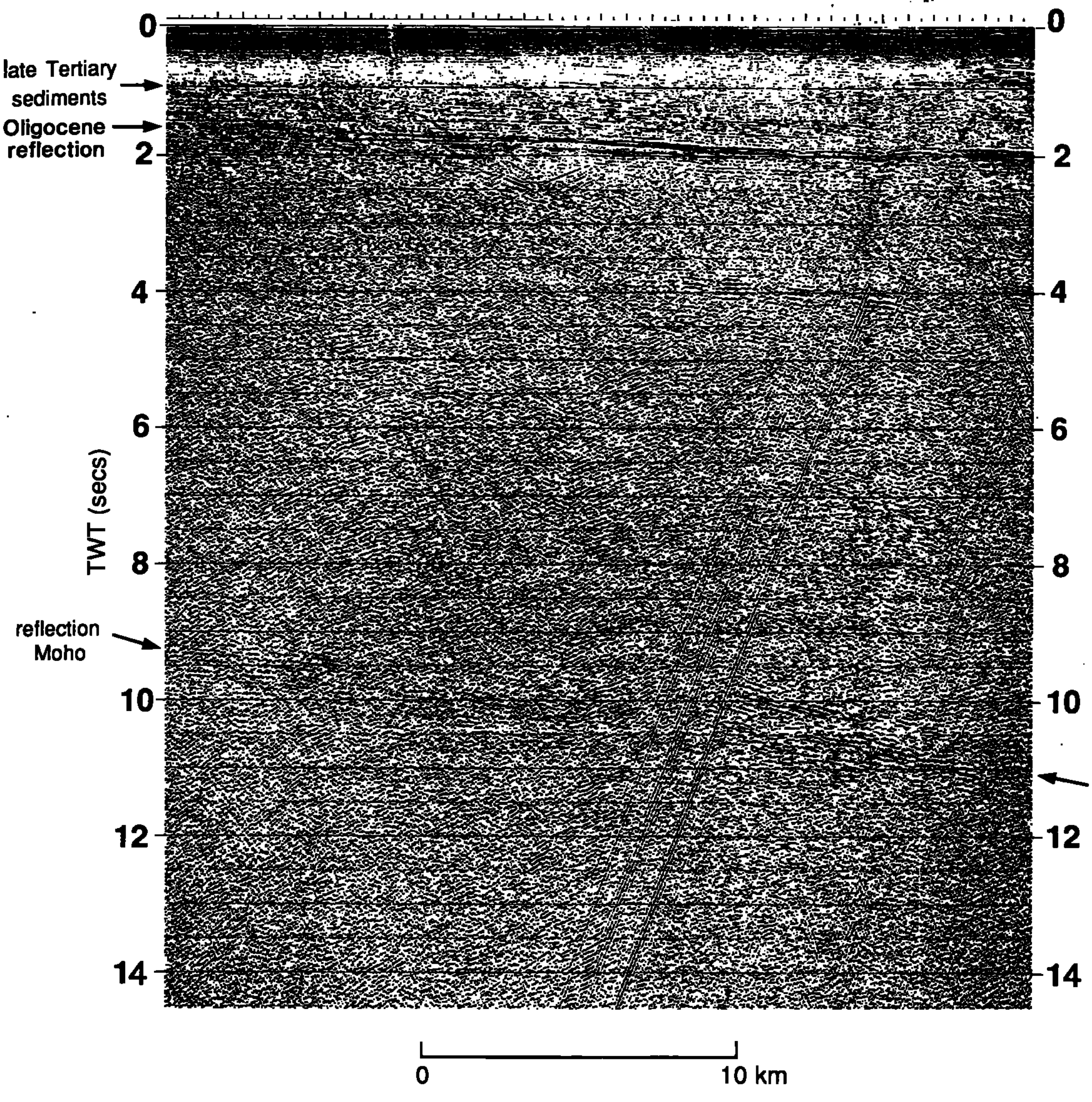


Fig. 3b: Part of the data from east of the Cape Egmont Fault Zone. Note the less intense lower crustal reflectivity, compared to that in Fig. 3a, and the eastward dipping reflection Moho. Note the crossing events between 2 and 3 s TWT, and just west of centre, that is interpreted as one of the many early-Tertiary half-grabens found within the Taranaki Basin.

FLEXURE OF THE LITHOSPHERE

Flexure of the lithosphere on a 100-500 km half-wavelength (the distance from the maximum deflection to the outer high (Fig. 6a)) is a well documented phenomenon both within the oceans (Bodine *et al.*, 1981) and the continents (Walcott, 1970). Flexure can be described mathematically as shown in the mechanical model of Fig. 6a. The wavelength and amplitude of flexural deformation have important thermal and geological implications. For example, the wavelength of flexure is thought to be a measure of the *effective thermal age* (time since the last major thermal or metamorphic event) of the lithosphere at the time it was loaded; the narrower the wavelength of flexure the younger the effective thermal age (Karner et al.,1983). For the South Taranaki Basin the half-wavelength of flexure is estimated from the palinspastic reconstructions of Pilaar and Wakefield (1978) and the deep seimic data to be about 170 km. This is a small flexural half-wavelength for continental lithosphere.

From the formulation of Watts and Talwani (1974) a flexural half-wavelength of 170 km for an assumed semi-infinite elastic plate, that has been buried by sediments, implies a flexural rigidity (D) for the plate of about 8 x 10^{22} N m. If this value of D is plotted on Karner *et al.'s* (1983) empirical plot of D Vs. effective thermal age at loading, an age at loading of about 45 Ma is estimated. Since loading by thrust sheets is inferred to have begun at about 20-30 Ma a true effective thermal age of the lithosphere, beneath the South Taranaki Basin, of about 70 Ma is implied. Thus even though the wavelength of flexure beneath the South Taranaki Basin is comparatively small for continental lithosphere, it is consistent with lithosphere that was thermally rejuvenated in the late-Cretaceous early-Paleogene by rifting and spreading

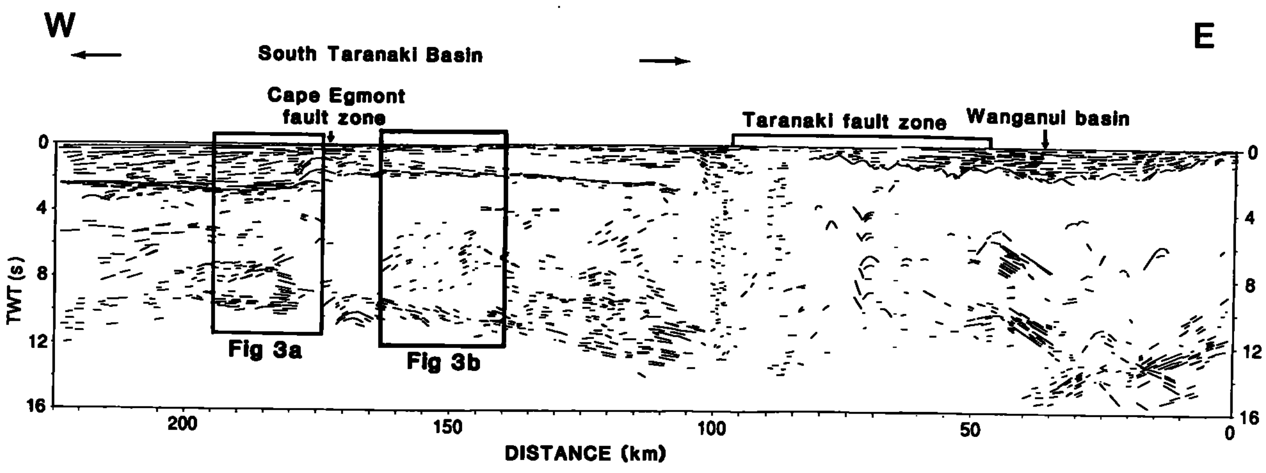
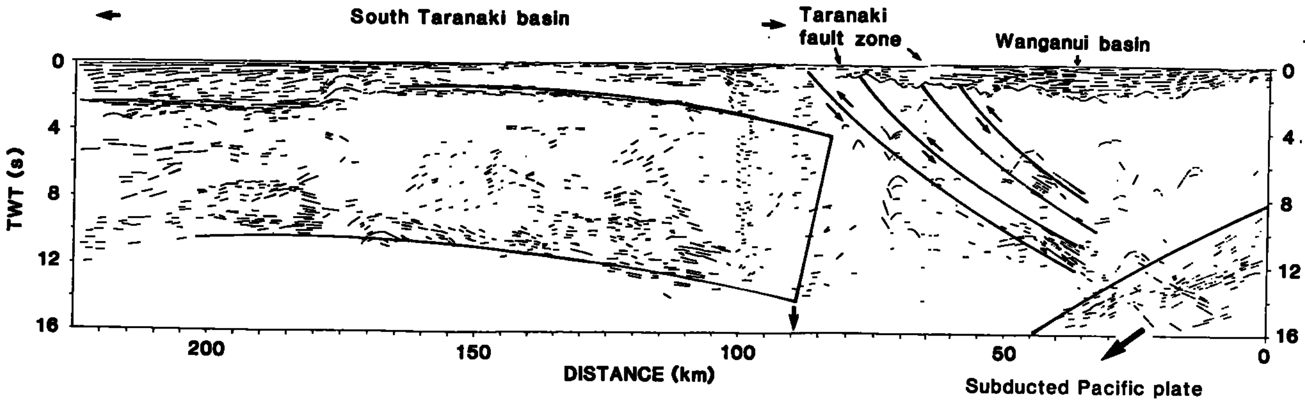


Fig. 4: A line drawing representation of the whole 220 km long deep seismic line. The localities for the data shown in Figs. 3a and 3b are indicated.



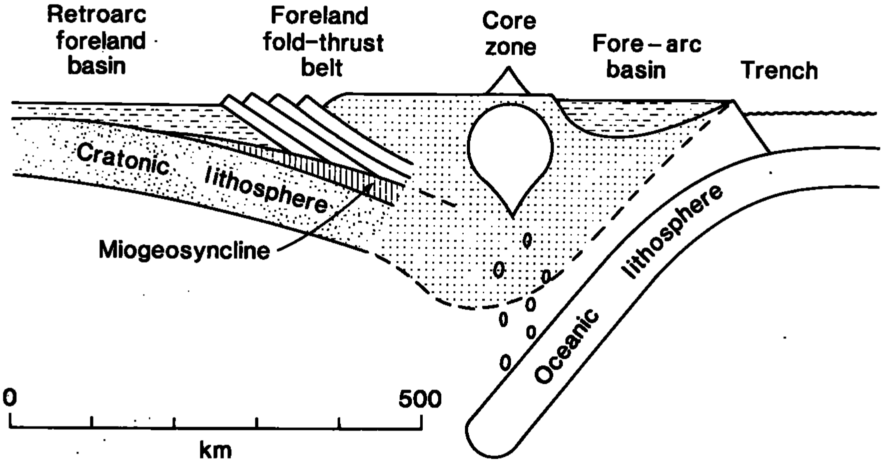


Fig. 5a: Line drawing of Fig. 4 with a tectonic interpretation superimposed.
Fig. 5b: A geological interpretation of the South Taranaki Basin in terms of Dickinson's (1974) and Beaumont's (1981) foreland basin model. Volcanism within the core zone may or may not be present.

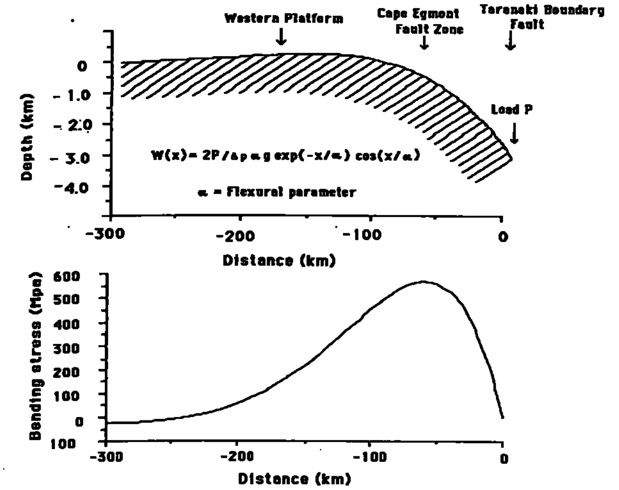


Fig. 6: A mechanical interpretation of flexure beneath the South Taranaki Basin. Flexure and associated bending stresses of an elastic plate with a free edge are described by the equations given by Watts and Talwani (1974). The equation for the flexural displacements, W(x), of an elastic plate that is loaded at its free-edge with a line load (P) is shown. For flexure associated with the South Taranaki Basin the outer high is estimated to be at about 170 km from the Taranaki Boundary Fault. If the fault is the locus of loading then an estimate for the flexural parameter (a) of 75 km and for the flexural rigidity (r) of about 8 x 1022 N m can be derived. In the model shown the load of the thrust sheets is taken as an equivalent line load of P = 1.5 x 1012 N.m and the density contrast (Δr) between material below and above the elastic plate is taken as 950 kg/m3. The load P is the "driving" load required to produce the 4 km (Fig. 2) of Oligocene-Miocene sediment amplified subsidence adjacent to the Taranaki Boundary Fault. Fig. 6b: Bending stresses calculated from the formulation of Watts and Talwani (1974). Note that a maximum occurs at a distance of about 70 km from the locus of the load. This maximum shows a correlation with the position of the Cape Egmont Fault Zone and therefore suggests a mechanical link between the Taranaki and Cape Egmont Fault zones.

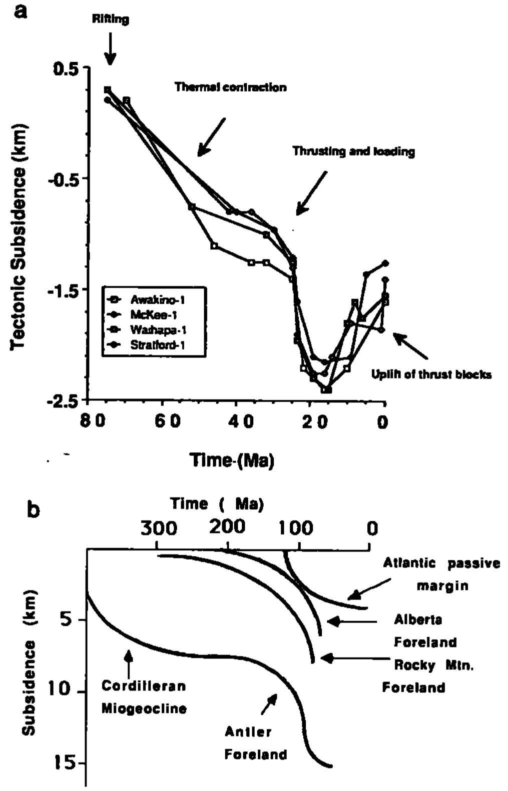


Fig. 7: A *stack* of 4 subsidence curves from the eastern Taranaki Basin. (Data from Hayward and Wood (1989)). These are back-stripped curves with a local Airy-isostatic model, and therefore represent an estimate of tectonic subsidence only. The locality of the drill holes are shown in Fig. 2. The interpreted phases of rifting, cooling, isostatic loading and uplift of thrust blocks are indicated. Fig. 7b: Schematic presentations of subsidence curves from sedimentary basins of North America (after Dickinson, 1976). Note how the exponential-decay shaped curves represent rifted margins, or miogeosynclines, and the exponential-increase shaped curves are associated with compressional or foreland basin structures. Together the total S-shape curve represents the complete foreland basin sequence.

associated with the opening of the Tasman Sea (Weissel *et al.*, 1977).

Another useful application of elastic flexure modelling is to check for spatial associations between major faults and the predicted location of maximum bending stresses due to flexure. For example, in Fig. 6b it can be seen that predicted bending stresses reach a local maximum 60-70 km west from the Taranaki Boundary Fault, where maximum curvature of the flexed plate occurs and in about the locality of the Cape Egmont Fault Zone. The implied bending stresses are of the order of 102 to 103 MPa, which is beyond most estimates for the maximum stress difference that the lithosphere can sustain (McNutt, 1980). Therefore, lithospheric failure by faulting is predicted. From this evidence, and the simple observation that the Taranaki and Cape Egmont Fault zones roughly parallel each other (Fig. 2), it is suggested that these two fault zones are mechanically linked by the process of induced bending stresses as shown in Fig. 6a and b. This link, however, may have been established as far back as the Mesozoic or Paleozoic when the original thrust and suture structure, now manifested most clearly by the Junction Magnetic anomaly (Hatherton, 1969), was put in place.

TIMING

Katz (1978) provided the first quantitative analysis of subsidence data from drillholes within the Taranaki Basin. He noted that, in general, subsidence and adjacent orogenic uplift are linked, although no specific mechanism was discussed. More recent and complete compilations of subsidence data from the Taranaki Basin, including curves showing subsidence with the local isostatic amplifying effect of sediment loading removed (*back-stripped* curves), have been made by Hayward and Wood (1989).

Four subsidence curves from just west of the Taranaki Boundary Fault are shown stacked together in Fig. 7a. From 80 Ma until about 35 Ma the curves show an exponential decay that is characteristic of passive margins where rifting is followed by thermal contraction (Sleep, 1971). At approximately 35 Ma another subsidence phase began slowly and then rapidly accelerated between 25 and 20 Ma. This is interpreted to reflect subsidence due to loading from a thrust front that initially formed east of the Taranaki Boundary Fault, and progressively advanced westwards, pushing a fore-deep trough in front of it.

The overall S-shape pattern of subsidence (Fig. 7a) is also recognised in North America (Dickinson, 1976). Fig. 7b shows representative forms of subsidence curves from North America that contrasts the exponential-decay shape of rift subsidence, with the exponential growth shape of compression induced subsidence. An S-shape curve therefore results from the full cycle of rifting followed by at least one lithospheric thermal- time constant (≈ 65 my) later by overthrusting.

In the last 15 my all of the curves in Fig. 7a show a strong uplift. This is explicable in terms of a uplift on a moving thrust front catching up with drill holes that are effectively fixed in the lithosphere. Furthermore, most of the drill holes are aimed at structural highs for exploration purposes, so that parts of the lithosphere that are not uplifted by the advancing thrust front tend not to be sampled.

PLATE TECTONIC CONSIDERATIONS

Plate tectonic reconstructions of Walcott (1984) indicate that the leading edge of the subducted Pacific plate, that is now beneath the South Taranaki Basin, first entered the trench off the East Coast of the North Island at 28 ± 8 Ma. This epoch corresponds closely with the phase of rapid subsidence seen in the drillholes (Fig. 7a). Therefore it is likely that it was thrusting associated with the initiation of a subduction zone, adjacent to the east coast of the North Island, that is at least partly responsible for the rapid subsidence phase in the period 20-30 Ma.

From paleobathymetry curves published by Hayward and Wood (1989) it appears that sedimentation did not initially keep pace with subsidence, and thus a 1-2 km trough developed west of the Taranaki Boundary Fault in late Oligocene time that was not completely filled until the mid-Miocene. An explanation for this lag in sedimentation is that as most of New Zealand was below sea level at this time, erosion and sedimentation would have been considerably muted. Only as the tempo of thrusting and compression increased in the early Miocene did the thrust sheets evidently develop sufficient topography to come above sea level and thus provide a ready source of terrigenous sediments.

That the New Zealand continent was mostly below sea level at this time is not surprising given the average crustal thickness, and thus the natural free-board, of the *New Zealand Continental Block* (Reilly, 1965). Seismic refraction experiments show that crustal thickness of the New Zealand continent, both on and off shore, for regions distant from the plate boundary, is about 25 km with extreme values of 20 and 29 km (Adams, 1962; Shor *et al.*,1971; Stern *et al.*, 1987); an average of 25 km will be adopted. From isostatic considerations we have:

$$\Delta Z_c (p_m - p_c) = \Delta Z_w (p_m - p_w)$$

where ΔZ_c is the difference in crustal thickness from an average value for continental crust at sea level, ΔZ_w is the water depth that will result from ΔZ_c and p_m, p_c, and p_w are densities of the displaced mantle, crustal rocks and water respectively. Taking the average crustal thickness of continental crust at sea level to be 30 km (Meissner, 1986) then ΔZ_c is -5 km. For values of p_m, p_c, and p_w of 3300, 2850 and 1000 kg/m^3, respectively, the free board (ΔZ_w) of the New Zealand continental crust, in the absence of any additional thermal or mechanical uplift forces, will be about -1000 m. Thus by the end of the Oligocene, about 35 Ma after the cessation of Tasman Sea spreading, the thermal effect of spreading would have decayed sufficently for most of New Zealand to be several hundred metres below sea level.

COMPARISON OF SOUTH TARANAKI BASIN WITH SOME EUROPEAN BASINS

In Fig. 8 a comparison is shown between deep seismic sections from the South Taranaki Basin, and two basins from within Europe; the Celtic Basin between the U.K. and Ireland and the Ebro (Spain) and the Aquitaine (France) basins found each side of the Pyrenees collision front. This comparison underscores the difference in deep structure between foreland basin and extensionally driven sedimentary basins demonstrated in Fig. 1. The Celtic Basin, that has an extensional origin, displays a flat Moho on the time section. In comparison, Moho reflections beneath both Taranaki and the foreland basins each side of the Pyrenees, are flexed strongly downwards.

The Pyrenees Collision Zone is different to the Taranaki Fault Zone in many regards. For example, there are several kilometres of topography created at the Pyrenees Collision Zone, whereas most of the surface expression of the Taranaki Boundary Fault is below sea level and in some places covered by up to 2 km of Plio-Pleistocene sediments. Also two foreland basin structures, the Ebro and the Aquitaine basins, have developed each side of the Pyrenees Collision Zone as thrusting emanated both north and south from the collision zone. Nevertheless, given these differences, there is a broad similarity in the deep structure of the South Taranaki Basin and the Ebro Basin on the south side of the Pyrenees Collision Zone (Fig. 8).

Estimates of shortening within the Pyrenees Collision Zone have been made, using geological mapping and palinspastic reconstructions, and are generally in the 100 km range (Williams and Fischer, 1984; Choukroune *et al.*,1989). Because most of the thrusts within the Taranaki Fault Zone are buried, an estimate for total shortening is difficult to make in the conventional way. An estimate can, nevertheless, be made with a comparison of the deep seismic sections (Fig. 8), and the mechanical and flexural parameters (Table 1), from the Taranaki Fault Zone and the Pyrenees Collision Zone. Similar amounts of crustal thickening, and therefore loading, have occurred at both zones. As loading is a function of the total shortening, it is likely that a similar amount of shortening (≈100 km) has occurred in both the Taranaki Fault Zone and the Pyrenees Collision Zone. However, it is possible that some, perhaps up to 50%, of the shortening in Taranaki Fault Zone, as manifested in crustal thickening, is a relic from thrusting in the Mesozoic.

IMPLICATIONS FOR THERMAL MATURATION AND FLUID MIGRATION

Recognition that the main subsidence phase of the south Taranaki was compressionally, rather than extensionally, driven has important thermal and hydrological implications.

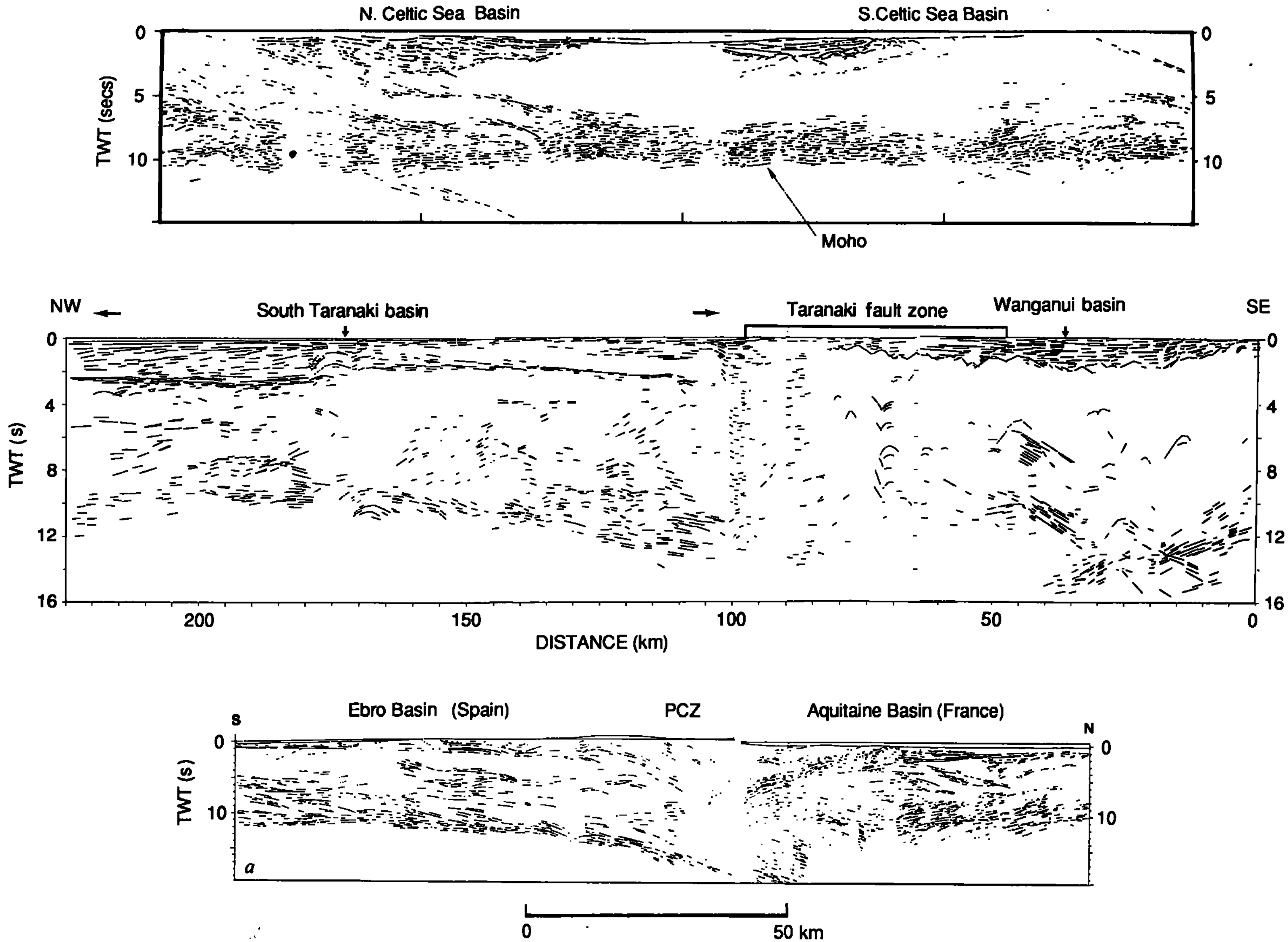


Fig. 8: A comparison of line drawing representations of deep seismic data from the Celtic Basin (Cheadle *et al.*, 1987), the South Taranaki Basin, and the Ebro and Aquitaine basins (Choukroune *et al.*,1989) found each side of the Pyrenees Collision Zone.

	TARANAKI	PYRENEES*
Flexural rigidity (D)	0.8×10^{23} N m.	1.5 x 1023 N m to 1.4 x 1021 N m. (i.e variable rigidity)
Depth of basin created by overthrusting and isostatic loading	≈ 3 km	≈ 5 km
Crustal thickening beneath overthrust	≈10 km	≈10 km
Required loading (P)	1.5 x 1012 N/m	1.6 x 1012 N/m
Amount of shortening	?	80-100 km

* Data from Brunet (1986).

Table 1. Comparison of flexural parameters between foreland basins formed at the Pyrenees Collision Zone and Taranaki Boundary Fault.

For example, the variation of heat flux with time due to the formation of extensional basins (McKenzie,1978) is straight forward, and predicts an initial heat flux peak followed by an exponential decay with a time constant of about 65 Ma (Sleep, 1971). Heat flux perturbations associated with overthrusting are, however, more subtle. Brewer (1981) shows that for regions overridden by thrust sheets of the order of 5-15 km thick, frictional heating will induce a positive heat flux anomaly that will persist for 5 Ma after thrusting has ceased. Thereafter a negative heat flux anomaly will be present and this will persist for 200-300 my.

Perhaps of even more import is the role of predicted fluid and hydrocarbon migration within foreland basins. Dickinson (1974), Oliver (1986), and Ge and Garven (1989) all argue that large scale compression and overthrusting can initiate transient fluid flow hundreds of kilometres into the foreland with consequences for hydrocarbon migration, sediment-hosted ore deposits and heat flux. For example, Oliver examines the spatial occurrence of oil and gas fields in the USA and notes that all major discoveries, with the exception of the Gulf Coast Province, are found in foreland basin structures behind orogenic belts. He argues that hydrocarbons have been pushed along porous paths within the foreland basins by advancing thrust fronts; the analogy Oliver uses is one of a tectonic *squeegee* driving fluids ahead of it. He also notes that meteoric water being driven through a basin will perturb geothermal gradients with the net effect of lowering the geothermal gradient near the thrust belt and raising the gradient far from the belt. Thus some forms of metamorphism and thermal maturation could be achieved far from the thrust front without the depth of burial that would otherwise be needed.

Within the Taranaki Basin all the present producing oil fields are found immediately adjacent or just west of the Taranaki Boundary Fault. Source materials drilled in the McKee, Kapuni and Maui fields are immature, suggesting that these accumulations have resulted from a migration of oil from a distant and deeper source (Palmer and Bulte,1989). If there has been a wide spread up-dip migration of hydrocarbons, driven by westward moving thrust sheets in a manner similar to that described by Dickinson (1974) and Oliver (1986), then promising prospects would be predicted within the vicinity of the Cape Egmont Fault Zone and even further westward out on the western platform.

CONCLUSIONS

This study demonstrates how deep seismic exploration has led to an interpretation of the South Taranaki Basin that was not immediately apparent from the approximate 25 000 km of shallow seismic reflection data shot in the 20 years prior to the deep survey. In particular, evidence from deep seismic and subsidence data point to a foreland basin style of late-Tertiary development for the South Taranaki Basin, rather than a rift graben as previously assumed.

Pronounced crustal thickening and a broad crustal flexure are the principal features seen in the deep seismic data. These features are most easily reconciled with crustal shortening and loading of thrust sheets within a 50 km wide Taranaki Fault Zone. Although the seismic resolution is at its poorest within the zone of the overthrusts, most of the crust is inferred to be involved in the overthrusting. On the basis of a comparison with a deep seismic section from the Pyrenees an estimate of about 100 km is made for shortening within the Taranaki Fault Zone. This is, however, an initial and tentative estimate.

A more general conclusion from this study is that deep seismic reflection studies can provide a new perspective on the geohistory and structure of sedimentary basins. For example, when we view a foreland basin on a 200 km-length and 40 km-depth scale we see that the dominant form of deformation is not faulting, but elastic flexure of the whole lithosphere. Faulting within the foreland basin is, to some extent, a secondary process.

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