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The dammed Hikurangi Trough: a channel-fed trench blocked by subducting seamounts and their wake avalanches (New Zealand–France GeodyNZ Project)

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ABSTRACT

The Hikurangi Trough, off eastern New Zealand, is at the southern end of the Tonga–Kermadec–Hikurangi subduction system, which merges into a zone of intracontinental transform. The trough is mainly a turbidite-filled structural trench but includes an oblique-collision, foredeep basin. Its northern end has a sharp boundary with the deep, sediment-starved, Kermadec Trench.

Swath-mapping, sampling and seismic surveys show modern sediment input is mainly via Kaikoura Canyon, which intercepts littoral drift at the southern, intracontinental apex of the trough, with minor input from seep gullies. Glacial age input was via many canyons and about an order of magnitude greater. Beyond a narrow, gravelly, intracontinental foredeep, the southern trench-basin is characterized by a channel meandering around the seaward edge of mainly Plio-Pleistocene, overbank deposits that reach 5 km in thickness. The aggrading channel has sandy turbidites, but low-backscatter, and long-wavelength bedforms indicating thick flows. Levées on both sides are capped by tangentially aligned mudwaves on the outsides of bends, indicating centrifugal overflow from heads of dense, fast-moving, autosuspension flows. The higher, left-bank levée also has levée-parallel mudwaves, indicating Coriolis and/or boundary currents effects on dilute flows or tail plumes.

In the northern trough, basin-fill is generally less than 2 km thick and includes widespread overbank turbidites, a massive, blocky, avalanche deposit and an extensive, buried, debris flow deposit. A line of low seamounts on the subducting plate acts as a dam preventing modern turbidity currents from reaching the Kermadec Trench. Major margin collapse probably occurred in the wake of a large subducting seamount; this seamount and its wake debris flow probably dammed the trench from 2 Ma to 0.5 Ma. Before this, similar dams may have re-routed turbidity currents across the plateau.

INTRODUCTION

Shallow Hikurangi Trough but deep Kermadec Trench – why?

The Hikurangi Trough, off eastern New Zealand, is at the southern extremity of the Tonga–Kermadec–Hikurangi subduction system (Lewis, 1980). It represents the topographic expression of oblique convergence between the over-riding Australian Plate and the subducting and underthrusting Pacific Plate (Fig. 1). The trough is largely a turbidite-filled structural trench but it merges southwards into a foredeep at the edge of a transform zone of highly oblique collision. The intracontinental transform zone links, via the Alpine Fault of South Island, to an opposite facing zone of oblique collision and subduction south-west of New Zealand. Relative motion between the plates, in areas remote from its deforming edge, ranges from 45 mm yr^{-1} in a direction of 266° at the trough's northern end to 38 mm yr^{-1} in a direction of 259° at its southern extremity (De Mets *et al.*, 1994) (Fig. 1). Obliquity, the angle between the direction of relative plate motion and the orthogonal to the deformation front, ranges from a low of 30° (nearly orthogonal) in the north to more than 80° (highly oblique) in the south.

The trough is shallow for what is largely a structural trench, at least partly because thickened oceanic crust of the Hikurangi Plateau is being subducted (Davy, 1992; Uruski & Wood, 1993; Mortimer & Parkinson, 1996). Its

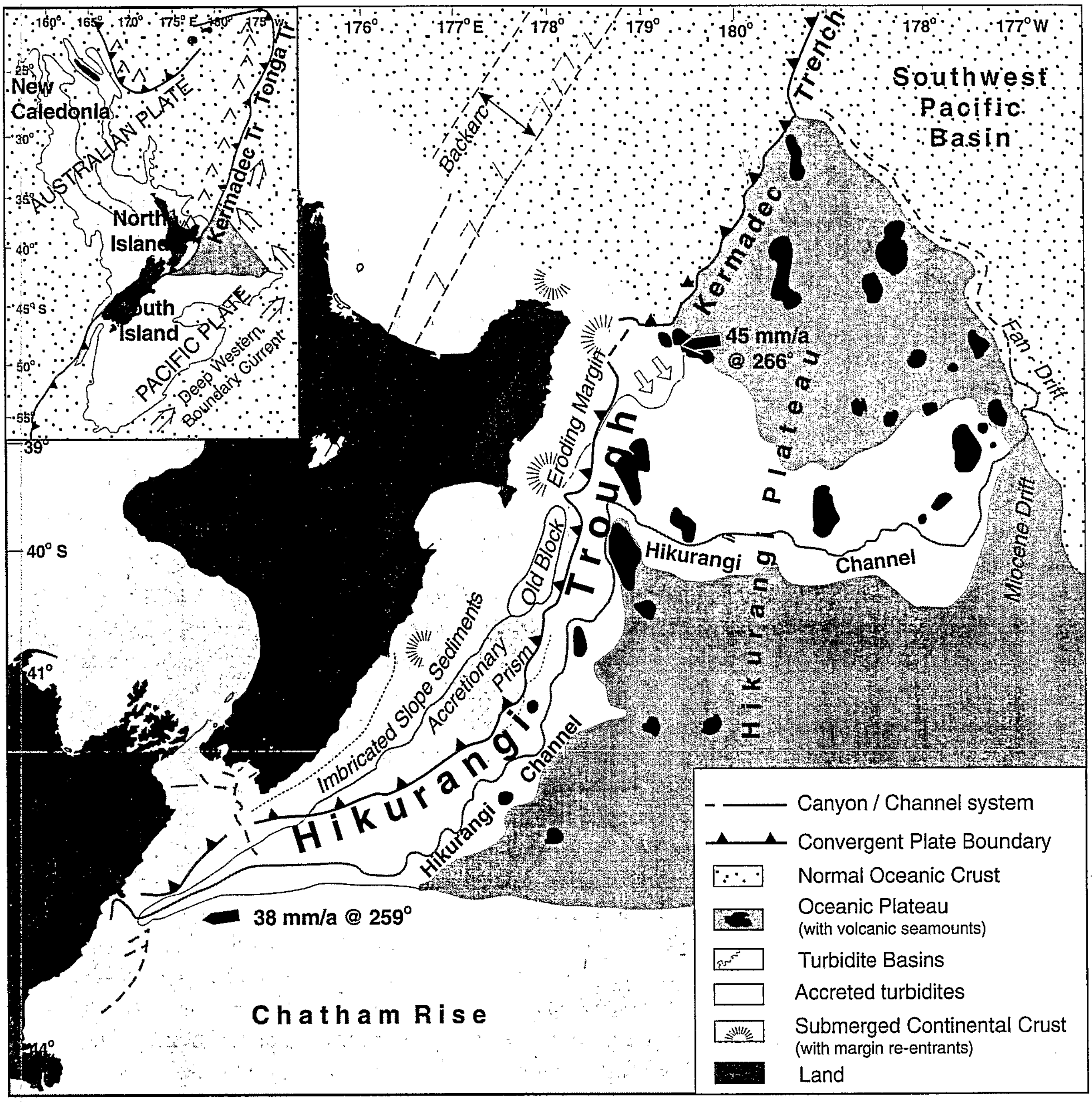
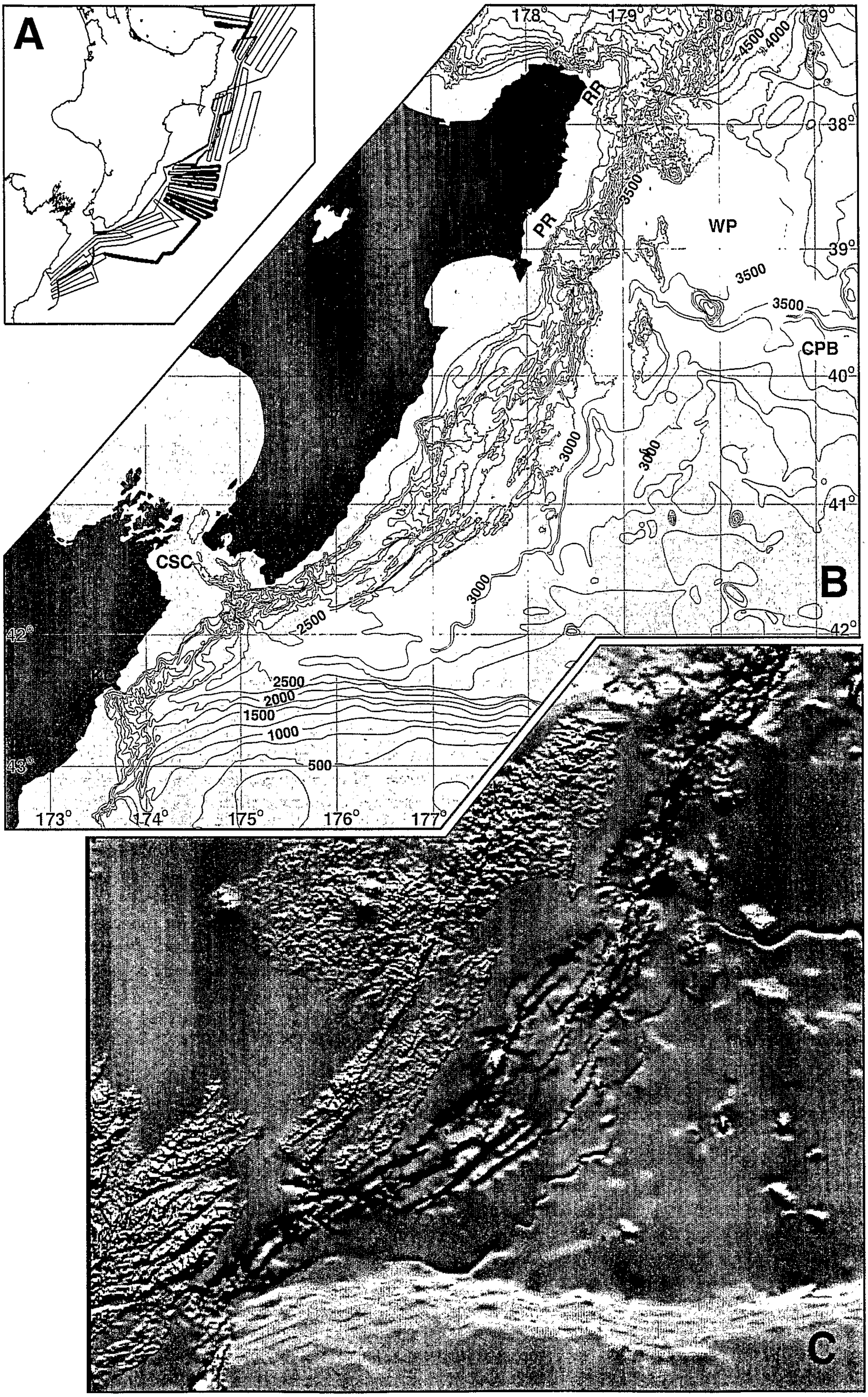


Fig. 1. Location of the Hikurangi Trough and main structural elements associated with it. Arrows indicate relative plate motion. Inset: plate boundary in the Southwest Pacific and route of Deep Western Boundary Current.

flat floor is about 720 km long by generally less than 70 km wide (Fig. 1), and it slopes from 2000 m to 2650 m deep in the 90 km long, intracontinental foredeep at the apex and from 2650 m to 3750 m in the main, subduction part of the trough (Fig. 2B). At the northern end, there is a drop of nearly 1 km into the head of the V-shaped southern Kermadec Trench, where Hikurangi Plateau is also being subducted; the change from shallow trough to much deeper trench does not coincide with any major change in the nature of the down-going plate.

The Hikurangi Trough is wholly on the down-warped edge of the Pacific Plate (Fig. 1). Its western boundary is mainly a curving, slope-toe, deformation front at the edge of the Australian Plate off eastern North Island

Fig. 2. Bathymetry of the Hikurangi Trough and adjacent areas. A: Tracks of GeodyNZ swath-mapping surveys, thin lines by *L'Atalante* EM12 and thicker lines by *Giljanes* HAWAII MR1. B: Bathymetry with contours at 250-m intervals; composite of swath surveys, archived and hydrographic soundings. Unshaded area is the Hikurangi Trough turbidite plain. RR = Ruatoria Re-entrant; PR = Poverty Re-entrant; WP = Whenuanuipapa Plain; CPB = Central Plateau Basin; CSC = Cook Strait Canyon; KC = Kaikoura Canyon. C: Shade-relief terrain model based on composite bathymetry with onshore relief, showing the trough with its axial channel system, accretionary ridges on the southern margin, and a blocky projection into the trough from the steep northern margin.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(Fig. 1). The trough's eastern flank is marked by a change to the undulating, seamount-studded topography of the Hikurangi Plateau (Wood & Davy, 1994) (Figs 1 and 2). At the trough's southern end, the subducting oceanic plateau merges into the continental Chatham Rise, which has been highly obliquely underthrusting beneath north-eastern South Island (Lewis *et al.*, 1986; Barnes, 1994; Barnes *et al.*, 1998), although motion may now be locked (Reyners *et al.*, 1997). The trough's northern limit, off East Cape (Fig. 1), is marked by a change from flat plain to the small, steep-sided, *en echelon*, enclosed basins of the southern Kermadec Trench (Collot *et al.*, 1996). In the same area, there is an apparent offset and change in trend of the plate boundary (Katz, 1974, 1982; Collot *et al.*, 1996). To the north, the axis of the Kermadec Trench slopes from over 4500 m adjacent to the trough, to 6000 m deep at the northern end of the Hikurangi Plateau, and from 7000 m to 10 000 m deep north of the plateau (Collot *et al.*, 1996; CANZ, 1997).

The evolution of the Hikurangi Trough and the cause of the discontinuity between the Hikurangi Trough and Kermadec Trench have long been causes of speculation (Brodie & Hatherton, 1958; Van Der Linden, 1969; Cole & Lewis, 1981; Katz, 1982; Davy, 1992). Many authors have considered that the apparent offset between trough and trench signifies a major structural break, related in some way to the NW/SE-trending continental edge of northern New Zealand with which it is aligned (Fig. 1). However, apparent continuity between an old structural trend on the over-riding plate and a major change in character on the downgoing plate is puzzling, particularly as convergence is oblique. The question of whether there is a major structural break between Kermadec Trench and Hikurangi Trough, or whether the discontinuity has some other cause, remains unresolved.

In this paper, we examine the main geomorphic features of the Hikurangi Trough and ask why it exists distinct from the Kermadec Trench.

Turbidite plain with a re-routed axial channel

The flat plain of the Hikurangi Trough is characterized by an active turbidite channel, the Hikurangi Channel (Lewis, 1994) (Figs 1 and 2). The Hikurangi Channel meanders three-quarters of the way along the trough from its southern end, before turning at right-angles out of the trough to deeply incise the Hikurangi Plateau before emerging at a distal fan at the edge of the Southwest Pacific Basin (Fig. 1). There, the Pacific Ocean's powerful Deep Western Boundary Current (Fig. 1 inset) sweeps along the scarp at the edge of the Hikurangi Plateau carrying sediment north-westwards into a turbidite–contourite 'fan drift' (Fig. 1). This deposit almost reaches the central Kermadec Trench (Carter & McCave, 1994; Collot *et al.*, 1996).

The puzzle is why channelled turbidity currents are routed out of the trough and across the plateau, rather than take the direct route to the central Kermadec Trench along the trough–trench axis.

Trench basins – a dynamic, interactive system

A filled trench is a highly dynamic type of sedimentary basin. It is being constantly consumed along one edge and compression in the adjacent margin implies a constantly varying input. There is on-going structural and sedimentary interaction between basin and flanking margin.

The nature of the adjacent Hikurangi margin changes significantly from south to north (Lewis & Pettinga, 1992; Collot *et al.*, 1996), and it is to be expected that the trough that interacts with it will also display significant lateral changes. In the southern intracontinental environment, a margin of slope clinoforms and turbidites is being sheared by transpressional plate boundary faults, mainly on the upper margin and adjacent land (Barnes & De Mercier Lépinay, 1997). Off southern North Island, there is a northward widening accretionary prism of offscraped Plio-Pleistocene trench sediments in front of an imbricate backstop of Cenozoic slope strata (Figs 1, 2) (Barnes & Mercier de Lépinay, 1997; Barnes *et al.*, 1998). The accretionary prism reaches a maximum width of over 70 km about half way along the Hikurangi Trough (Lewis, 1980; 1986b; Davey *et al.*, 1986a; Lewis & Pettinga, 1992). Further north, the accretionary prism narrows to less than 10 km wide in front of a steep block of relatively old, Cretaceous rocks (Lewis, 1985; Collot *et al.*, 1996; Lewis *et al.*, 1997). At the northern end of the trough, the margin is steep, without an accretionary prism, and is inferred to have been tectonically eroded by a seamount-studded downgoing plate (Collot *et al.*, 1996).

In this paper, we examine the lateral changes of the trough and the evolutionary changes recorded in its fill and relate these to interactive lateral and temporal changes in the adjacent margin. In this way, we consider that the Hikurangi Trough may provide models for many of the complexites of highly deformed, trench-fill basins preserved in fold belts around the world (Underwood & Bachman, 1982), including the adjacent mountain ranges of New Zealand (MacKinnon, 1983; Barnes, 1988; Barnes & Korsch, 1990).

METHODS

The primary dataset for this study is the first extensive swath mapping survey undertaken in New Zealand waters. It is part of the GeodyNZ Project undertaken jointly by New Zealand and French institutions to study key offshore segments of the obliquely compressive plate boundary through the New Zealand region (Collot *et al.*, 1994, 1995). Most of the data were collected in December 1993 on the French Research Vessel *L'Atalante* using a hull-mounted, Simrad EM12D, dual, multibeam, swath-mapping system. The EM12D measures depth and seabed reflectivity from 162 beams that cover a swath that is about seven times the water depth. Post-processing produces maps of both contoured bathymetry and seabed reflectivity from a swath of seabed that is generally 12–22 km wide in the study area. In some circumstances, 12 kHz sound may penetrate soft sediment and be reflected from strongly reflective layers that are buried up to several metres. Simultaneously, *L'Atalante*'s six-channel, twin 75-in^3 airgun system, its Barringer M-244 proton magnetometer and its Brodenseewerk KSS30 gravimeter were collecting seismic reflection, magnetic and gravity data along the same, mainly slope-parallel tracks (Fig. 2A).

Additional swath bathymetry and swath imagery was obtained in August 1994 using an MR1 system chartered from the University of Hawaii and towed from the New Zealand vessel *Giljanes* mainly along tracks transverse to the margin (Fig. 2A). *L'Atalante*'s seismic data were linked using a single channel 120-in^3 airgun seismic system and 'groundtruthed' with dredges and a piston corer. The new information was augmented by data collected from earlier surveys (Lewis & Pettinga, 1992; Lewis, 1994).

Preliminary processing of swath datasets included correction for velocity of sound using sea-water column velocity profiles obtained from expendable bathythermographs. Cores and rock samples were described and representative samples were analysed using standard settling tube and pipette methods. Estimates of nannofossil age have been supplied by Stratigraphic Solutions of Lower Hutt and ash correlations by Victoria University of Wellington.

RESULTS: IMAGES OF A DAMMED TRENCH

New levels of detail and new insights

The new swath bathymetry, swath imagery, geophysical data and samples, when merged with archived datasets, show for the first time the detailed shape, trends and structure of different segments of the Hikurangi Trough (Fig. 2). They show the longitudinal changing shape and bedforms of the trough's axial channel. They show the shape and distribution of small and large seamounts that constrict and even dam the trough. They show an enormous, blocky projection into the northern trough that is interpreted as a major avalanche deposit and, in front of this, a broad plain above a trough-filling debris flow deposit. The morphology and structure can be interpreted in terms of the nature and timing of events that have controlled the evolution of the trough as a major sedimentary basin. The datasets used in this study include new and detailed information from the adjacent continental slope and these parts are used elsewhere in structural interpretations of the margin (Collot, 1995; Collot *et al.*, 1995, 1996; Barnes & Mercier de Lépinay, 1997; Barnes *et al.*, 1998).

Active and inactive apical canyons

At the south-eastern end of the Hikurangi Trough, there are several dendritic canyon systems that are inferred to have been major conduits for sediment input (Lewis, 1994). The Kaikoura Canyon, which is the only canyon to presently intersect a longshore sediment transport system, has been identified as the principal conduit for sediment-laden flows during the Holocene (Carter *et al.*, 1982). Its position close to major transpressional plate-boundary faults (Fig. 1 and inset) (Barnes et al., 1998) means that it is subject to frequent, severe earthquake stress (Van Dissen, 1991).

The canyon head is within a few hundred metres of the shore platform (Fig. 3) and in water less than 20 m deep (Lewis *et al.*, 1998). It intercepts about 1.5×10^6 m^3 yr^{-1} (1.5 km^3 kyr^{-1}) of the main, northward-moving, nearshore, sediment transport system, which is fed by large rivers draining the mountains ranges of South Island (Lewis & Barnes, 1998). The canyon's headwall slopes at up to 30°, and the canyon axis slopes at 1–2° from 1000 m deep to the narrow, southern extremity of the Hikurangi Trough at 2000 m deep (Fig. 3A). The lower canyon has strong acoustic backscatter (Fig. 3B), commensurate with Holocene and Late Pleistocene gravel turbidites that have been cored from the lower canyon (Fig. 3A) (Carter *et al.*, 1982; Lewis & Barnes, 1998). In contrast, the Pegasus Canyon, which enters the head of the Hikurangi Trough from the south (Fig. 3A), is weakly backscattering, corroborating an inference that this canyon is mud filled and has probably been inactive since late in the last glacial age (Herzer, 1979).

Immediately north of Kaikoura Canyon, three of the six small Kowhai Canyons are strongly backscattering, implying that these too have coarse sediment in their axis (Fig. 3B). Each of these canyons is formed on the mid-slope from the junction of several, small gullies, spaced generally 1–2 km apart along the nearly straight upper slope and shelf edge. The outer shelf here is over 6 km from land, over 100 m deep, and well beyond major input from the modern, nearshore sediment prism. In the absence of cores and of a modern nearshore sediment supply, the strong backscatter is inferred to be coarse sediment that is either relict or derived from slope failure in the gully heads. In either case, the contribution of these small canyons to the Hikurangi Trough's sediment budget is likely to be minimal, compared with input from the major, nearshore sediment transport system that feeds the Kaikoura Canyon.

The intracontinental head of the trough

Strongly backscattering areas from the Kaikoura Canyon and Kowhai Canyons coalesce in the head of the Hikurangi Trough (Fig. 3B). From here almost to the confluence with the Cook Strait Canyon system, a 6–10-km-wide, flat-floored trough separates deformed

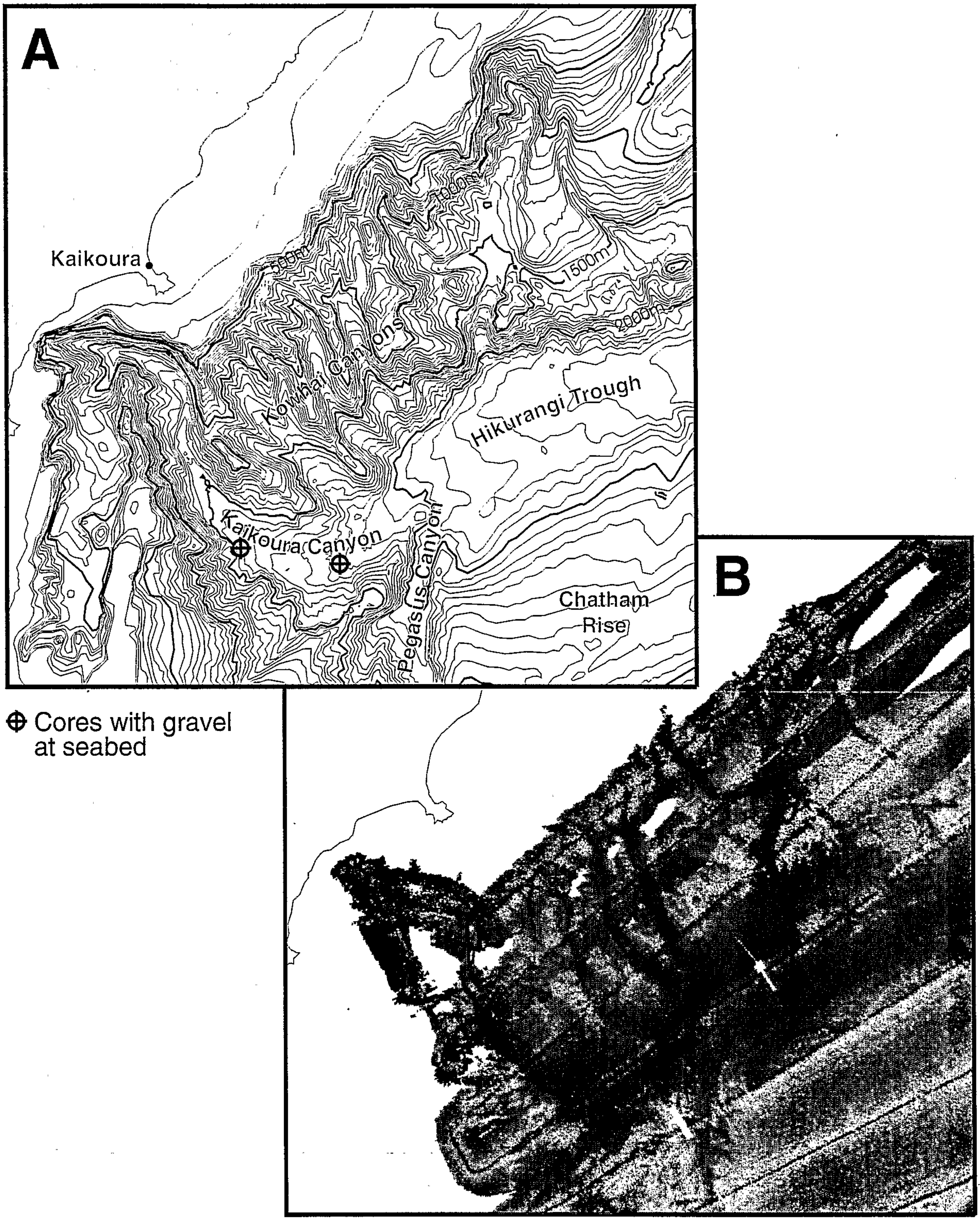


Fig. 3. Bathymetry and sonar backscatter of the southern extremity of the Hikurangi Trough and of its feeder canyons. A: Bathymetry with contours at 50-m intervals showing the Kaikoura Canyon's head within a few hundred metres of the shore. B: Sea-floor backscatter, where dark is strong backscatter from the southern extremity of the Hikurangi Trough and from the lower part of Kaikoura Canyon and some of the Kowhai Canyons.

slope sediments on the left bank from the flank of the Chatham Rise to the right (Figs 4A and 5) (Lewis *et al.*, 1986). The trough here occupies a foredeep between obliquely underthrusting Chatham Rise and the obliquely overthrusting transpressive ranges north of Kaikoura (Fig. 1) (Collot *et al.*, 1996; Barnes *et al.*, 1998). On the left bank, slope sediments have been elevated, perhaps by compression seaward of the main transpressional front, into a terrace that is 500 m above the trough (Fig. 4), and the scarp between terrace and trough is heavily

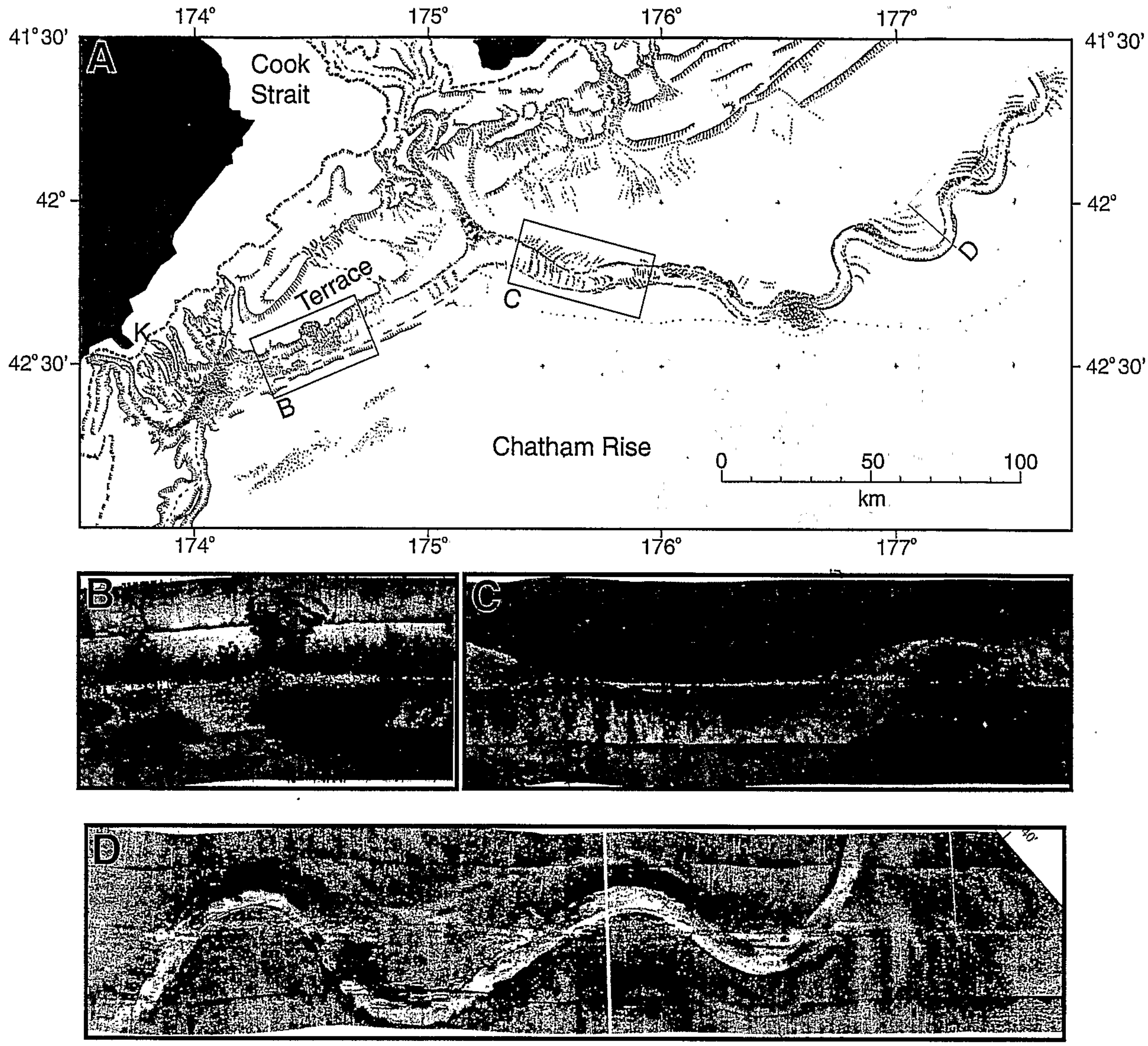


Fig. 4. Backscatter in the southern Hikurangi Trough. A: Diagrammatic representation of topography and backscatter, where ticked lines are scarps, ticked broken line is the shelf break and stippling represents strong reflectivity from channel and overbank sediment waves. Rectangles are locations of MR1 images shown in B–D. B: Backscatter from the intracontinental head of the Hikurangi Trough showing regularly spaced slope gullies and slope failure with dark, debris flow deposits in the trough. C: Weak backscatter from the narrowing and deepening channel floor in southern subduction trench showing low, long-wavelength sediment-waves in the channel axis, on the flanks of the left-bank levée, and on the insides of bends. D: Weak backscatter from the 230–280-m-deep, meandering channel floor in the southern Hikurangi Trough showing mud-wave trains tangential to overflow at bends on both sides, mud-waves subparallel to the channel on higher left bank only.

incised by gullies and small avalanche scars (Fig. 4A,B). The small gullies are generally 1–2 km apart. In the trough below the avalanche scars, speckled, strongly backscattering sediments indicate the extent of blocky avalanche deposits derived from the scars. Backscatter from trough sediments decreases north-eastwards from being more reflective than the adjacent slopes in the south to being generally less reflective near Cook Strait. At the edge of the intracontinental zone foredeep, just before the confluence with the Cook Strait Canyon, there are transverse bedforms with a wavelength of 3–4 km in a 5-km wide channel, the Hikurangi Channel (Fig. 4A), that is incised up to 80 m below the general level of the rapidly widening trough (Lewis, 1994). The transverse waves are about 10 m high and are more strongly back-scattering on their downslope face.

The wide, southern oblique subduction trough

Off southern North Island, the Hikurangi Trough extends as a flat, generally 50–80-km-wide, turbidite basin, in a gentle curve seaward of an accretionary prism of offscraped trench sediments that reaches a maximum width of about 70 km at about 41°S. (Figs 1, 2 and 5; Lewis, 1980; Davey *et al.*, 1986a,b; Barnes & Mercier de Lépinay, 1997). New and archived seismic data show parallel-bedded, turbidite basin-fill overlying the down-

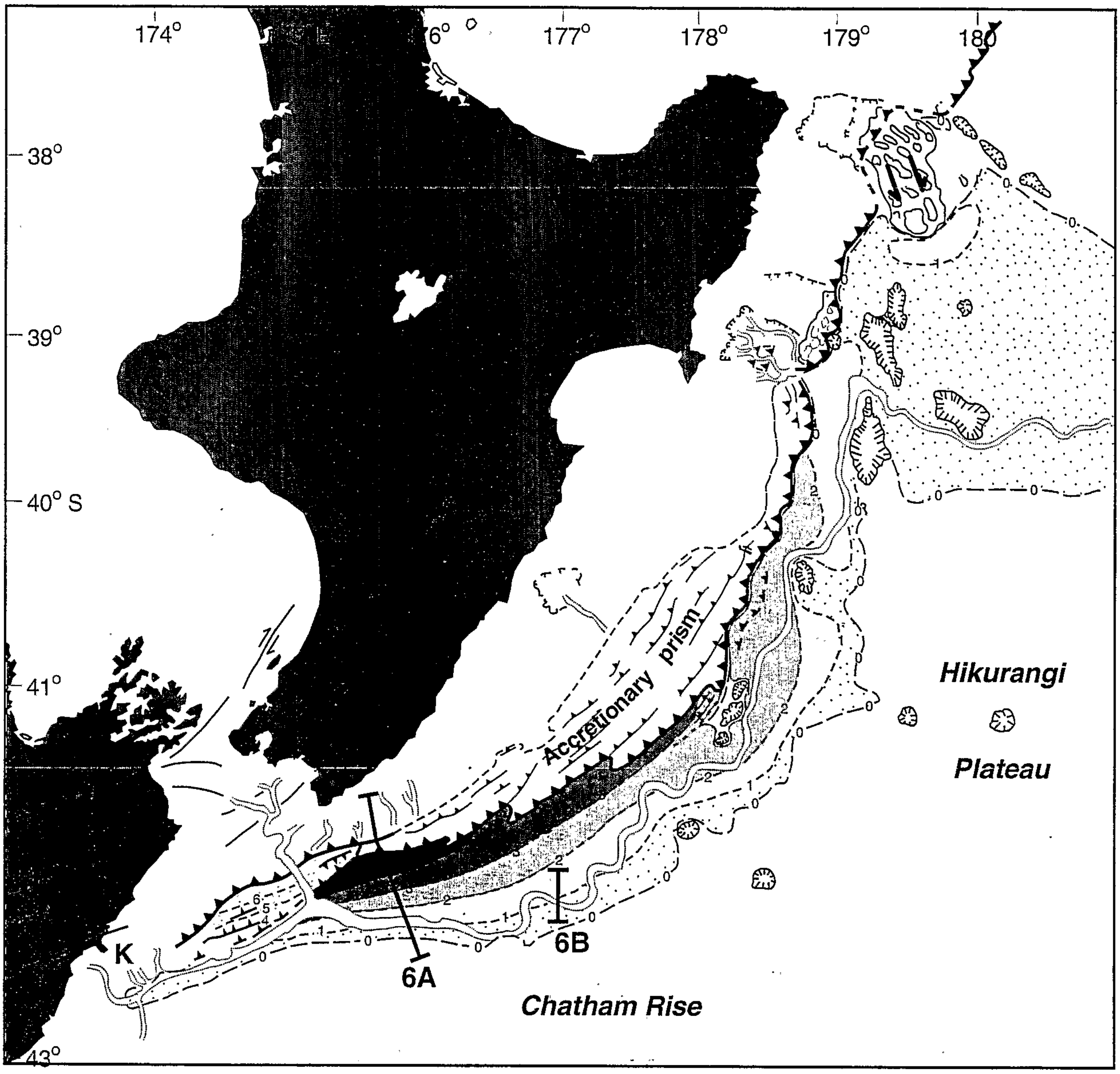


Fig. 5. Generalized thickness of turbidite-fill in kilometres, compiled from published, archived and new seismic data. Note that this does *not* include the thickness of Hikurangi Plateau pelagic drape. In the southernmost trough, sediments in front of the main deformation front (thick flagged line) include slope sediments. On the margin, a broken line represents the approximate limit of accreted trench-fill and transverse lines show major canyons. Also shown are the locations of profiles shown in Fig. 6.

warped northern flank of the Chatham Rise (Figs 5 and 6A). The fill thickens towards the deformation front, where it is about 5 km (over 4 s TWT) thick in the south, thinning to about 2.5 km thick at 41°S. Note that this is the thickness of parallel-bedded turbidites only, *not* the total sediment thickness. There are thick sedimentary sequences beneath the terrace to the south of Cook Strait, but these are believed to include slope sediments, older rise sediments, as well as trough turbidites (Fig. 5; Henry *et al.*, 1995). The southern trough is characterized by a deeply incised, meandering channel (Fig. 4) with well-developed levées and mud-waves (Figs 4D and 6). The flat channel floor is conspicuous on swath images because it has much weaker backscatter than the channel walls, levées and overbank plain (Fig. 4C,D). This contrasts with the much stronger reflectivity on seismic records (Fig. 6B). Reasons for these apparent anomalies are discussed later.

Beginning at the confluence with the large Cook Strait Canyon system (Figs 2 and 4A), swath imagery shows moderately strong backscatter and irregular bedforms 2.5 km in wavelength in the lower end of the canyon (Fig. 4A). A core from the canyon axis contains pebbly layers (Pantin, 1972). However, it is known that the two main arms of the Cook Strait Canyon are blocked by slumps (Carter, 1992), and that the only unblocked arm incises the outer shelf remote from any significant, modern, sediment transport system. Thus, it is inferred that the Cook Strait Canyon system is unlikely to have been an abundant supplier of sediment to the Hikurangi

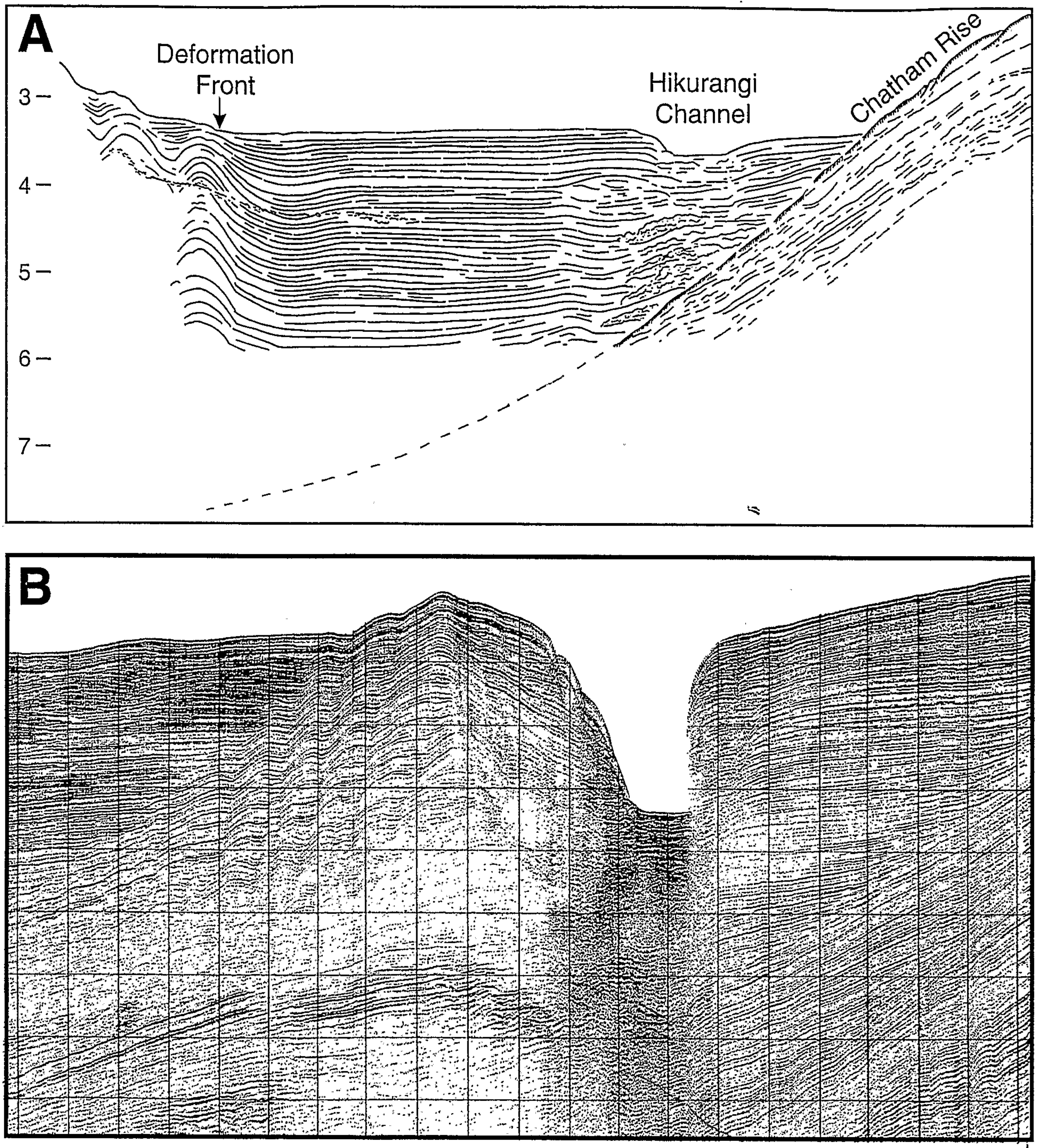


Fig. 6. Airgun seismic profiles of the southern Hikurangi Trough showing channel, levée and overbank deposits. A: Tracing of low-frequency profile showing Plio-Pleistocene turbidites onlapping the Chatham Rise and thickening towards the deformation front with limited lateral migration of the channel in aggrading fill. Scale at left in seconds of two-way travel time, roughly equivalent to kilometres in turbidite-fill. Vertical exaggeration about 10 : 1 (from Mobil 72-119). B: High-frequency profile showing high reflectivity from aggrading channel axis that has little lateral migration with time. Horizontal scale lines at 0.1 s (75 m in water or up to 100 m in sediment). Vertical lines are about 1.6 km apart. Vertical exaggeration about 25 : 1.

Trough since the last post-glacial rise of sea-level. Several small canyons on either side of Cook Strait Canyon also have moderately strong backscatter (Fig. 4A) and contain sandy layers (Pantin, 1972) but, like the Kowhai Canyons to the south, they begin either at the outermost shelf or as small, regularly spaced, 'headless' gullies on the seaward flanks of midslope ridges (Fig. 4A). Without an abundant sediment supply from a modern coastal sediment transport system, these canyons are also unlikely to be major suppliers of sediment to the trough.

From the confluence with the Cook Strait Canyon, the Hikurangi Trough widens rapidly to over 80 km wide and the Hikurangi Channel extends eastwards for about 120 km, hugging the northern flank of the Chatham Rise. The channel floor narrows from over 11 km wide to less than 3 km wide and deepens from about 210 m to 280 m below the left-bank levée, which is generally 30–100 m above the adjacent turbidite plain. Backscatter images show conspicuous bedforms in the channel axis (Fig. 4A,C). Close to Cook Strait, light and dark banding indicates transverse waveforms with a wavelength of 2.5 km and superimposed small waveforms with a wavelength of about 600 m (Fig. 4A,C). Bedforms continue from the weakly backscattering channel floor onto the more strongly backscattering flank of the left-bank levée (Fig. 4C), indicating a considerable thickness of flow capable of forming such bedforms. In the narrower and deeper part downstream, channel bedforms have a wavelength of about 400 m and vertical relief of the bedforms is small; contrasting backscatter represents sediment contrasts rather than relief. A core from the channel axis in this area contains a micaceous, medium to coarse sand turbidite at least 1.75 m thick beneath a 0.15-m-thick hemipelagic mud layer (Pantin, 1972). The edges of the channel, particularly on the outsides of meanders, are sharply imaged, indicating steep walls, whereas the insides of bends are diffuse, with transverse waveforms suggesting an analogy with fluvial point bars (Fig. 4C). At two places, there has been significant collapse of the channel walls (Fig. 4A). The smaller of the two involves about 6 km of the left bank and extends only half way across the channel floor. The larger occurs where the channel strongly undercuts the Chatham Rise slope (Barnes, 1992), and involves over 10 km of channel wall, with debris apparently extending almost across the channel axis. Beyond the high left-bank levée, the turbidite plain slopes down towards the slope-toe deformation front, where faint bedforms transverse to the front suggest secondary flow along the toe of the slope with perhaps faint, fan-like structures off the ends of some small canyons (Fig. 4A).

After hugging the Chatham Rise as far as it can, the channel meanders northwards along the outer edge of the turbidite-filled basin. Its weakly reflective floor narrows gradually to less than 2 km wide, being incised 240–280 m below the higher left-bank levée and 140–190 m below the right-bank levée. The channel forms a continuous series of meanders with a radius of 6–8 km (Figs 2B, 4). At the outside of each bend and continuing downstream from its centre, the channel walls are strongly reflective, steep, and their tops are scalloped suggesting collapse of 2–3 km wide wall segments (Fig. 4D). These features imply erosion by currents strong enough to encourage channel migration but nowhere are there any abandoned meander segments. In the same area, seismic profiles show mud waves 5–15 m high and 2–3 km in wavelength migrating up the levée backslopes on both banks (Fig. 6B) (Lewis, 1994). Swath images show two trends of the mud waves (Fig. 4A,D). One trend, which occurs on both sides of the channel, is subparallel to the strongly reflective outside of each meander bend and in a train aligned tangentially to the bend. This form is consistent with centrifugal overflow from the channel. The other trend, which occurs only on the higher left bank levée, is generally fainter and subparallel with the levée crest, being best displayed on the insides of bends between the well-developed waves on the outside of bends. This form may be a product of the same Coriolis-influenced overflows that elevated the left bank levée.

At 41°05′S the channel touches the seaward side of a small seamount (Figs 5 and 7) close to the plate boundary (Davey *et al.*, 1986a,b). Immediately to the south, the left-bank levée is low, perhaps permiting any secondary flow along the toe of the slope to return to the channel. Here, trench-fill up to about 3 km (2.5 s) thick is being rucked into numerous small 'protothrusts' between the knoll and the first significant thrust-bounded ridge at the toe of the slope (Fig. 7A). It is inferred that the deformation front is in the processes of re-establishing up to 24 km in front of the pre-existing one but it is unclear whether there is yet any decollément beneath the protothrusts (Barnes *et al.*, 1998). The 'protothrusts' in trough sediments are 15–30, mainly landward-dipping, planes of low reflectivity with bedding plane displacements that are estimated to range up to 25 m (Lewis & Pettinga, 1992). The low reflectivity of the protothrusts suggests alteration of the sediment's physical properties, perhaps by the passage of fluids from overpressured layers to landward. Swath images show that protothrusts are aligned subparallel to the deformation front and that they are laterally continuous for 5–15 km (Fig. 7B,C). The 'protothrusts' record the first stage in the tectonic loss of sediments from the basin.

Mahia depocentre and margin re-entrant

North of a small seamount at 40°15′S, the trough opens out into a 3400-m-deep plain that is about 150 km long by 20–30 km wide. It is sandwiched between the toe of the margin and several large seamounts that include Mahia Seamount and Gisborne Knolls (Fig. 8A). The plain is the surface expression of a basin depocentre which is generally 1.5–2.0 km (1.2–1.8 s) thick at the deformation front (Figs 5 and 8B). The adjacent continental margin is steep and has either a narrow, and locally blocky, accretionary prism or no accretionary prism at all (Collot *et al.*, 1996; Lewis *et al.*, 1997).

As the axial channel enters this plain, there are sediment waves 2–2.5 km in wavelength and up to 8 km long aligned transverse to the probable direction of centrifugal overbank flow from the outside of the bend (Fig. 8A). To the north, the channel again hugs the outer edge of the trough, constrained partly by the western edge of Mahia Seamount. At the northern end of Mahia Seamount, the floor of the channel changes from having lower backscatter than the surrounding plain to having higher backscatter, and the banks become lower, more gently sloping and without well-developed levées. The loss of a left-bank levée may allow secondary sheet flows on the plain to enter, or re-enter, the channel. Between Mahia Seamounts and Gisborne Knolls, where the channel has meandered 800 km from its source at Kaikoura, it turns at right-angles out of the structural trench,

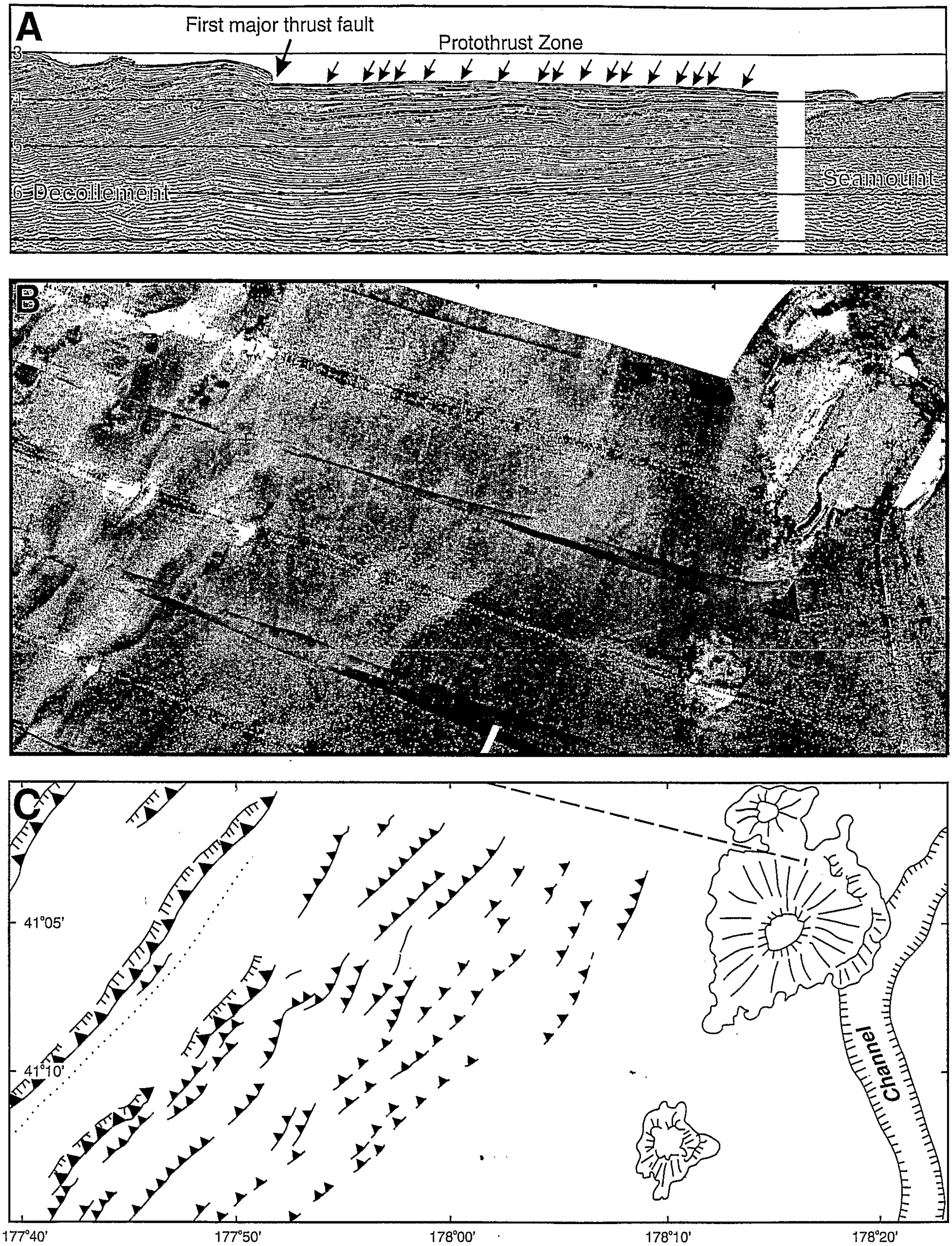
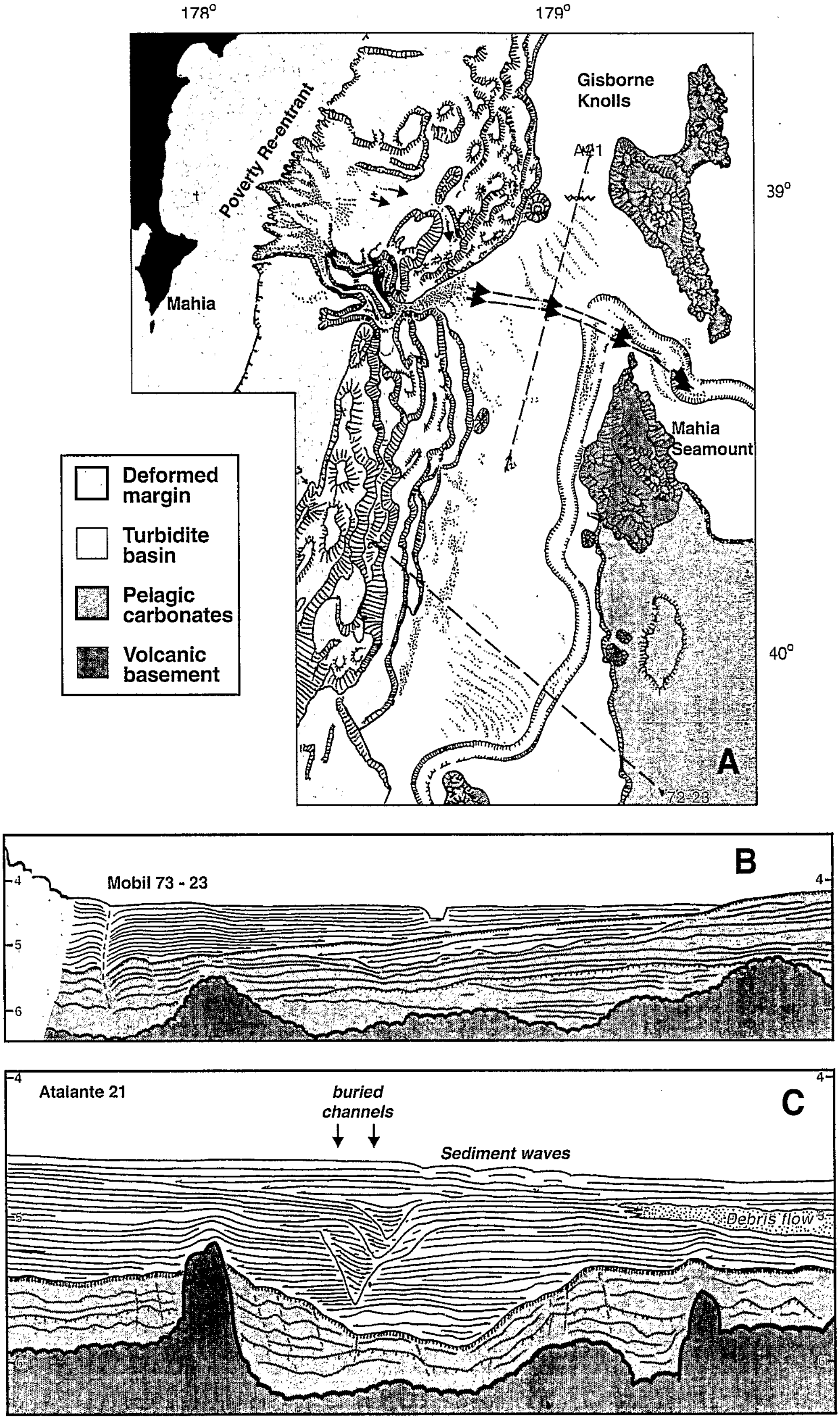


Fig. 7. Initial deformation at the western edge of the Hikurangi Trough. Protothrusts extend 15–25 km in front of the primary thrust fault at the toe of the slope; at this locality there is a small seamount near their seaward limit. A: Seismic profiles of protothrusts seen as mainly landward-dipping planes of low reflectivity with small (less than 25 m) offsets. Arrows indicate the more conspicuous protothrusts (from Davey *et al.*, 1986a,b). B: Mosaic of MR1 swath backscatter images showing NE–SW trend of protothrusts, which are laterally continuous for 5–15 km. C: Structural interpretation based on seismic profile and swath reflectivity; small flagged lines are protothrusts, large flagged lines are main thrust faults at base of scarps; trough seamounts and axial channel are also shown. Broken line is position of seismic profile.


178°
179°
39°
40°
Gisborne
Knolls
Poverty Re-entrant
Mahia
Mahia
Seamount
A41
72-23
A
Deformed
margin
Turbidite
basin
Pelagic
carbonates
Volcanic
basement
B
Mobil 73 - 23
4
5
6
C
Atalante 21
buried
channels
Sediment waves
Debris flow
4
5
6

exhibiting 'point bar' deposition at bends, and heads eastwards across the Hikurangi Plateau, before debouching onto a fan-drft at the edge of the Southwest Pacific Basin (Lewis, 1994).

The latitude at which the channel turns right coincides with a significant re-entrant in the continental margin (Fig. 8A), which has been inferred to be primarily a scar produced by the impact of a subducting seamount or seamounts against the soft sedimentary rocks of the margin (Lewis & Pettinga, 1992; Collot *et al.*, 1996). Incised into the re-entrant is the Poverty Canyon system. This is the only significant canyon system to reach the trough for nearly 300 km, canyons to the south being trapped in a wide baffle of accretionary slope basins (Lewis & Kohn, 1973). The Poverty Canyons incise only the outer shelf at depths greater than 80 m deep, so they do not trap the nearshore sediment drift during interglacial periods. However, regularly spaced gullies and tributary canyons, which incise the linear shelf break for 40 km, all have high backscatter in their axes, suggesting coarse sediment input, perhaps from headward sapping. Several canyons merge into a main Poverty Canyon, which emerges into the trough in the centre of a deep V-shaped re-entrant in the deformation front.

The V-shaped re-entrant defines a small, triangular, enclosed depression, which is separated from the main part of the trough by a low rise aligned more or less with the deformation front to north and south. There is no indication in seabed bathymetry or imagery of a modern connection between the Poverty Canyon and the Hikurangi Channel, which is only 30 km away. However, seismic profiles show infilled channels, with axes buried 0.4–1.0 km (0.4–0.9 s TWT) beneath the trough (Fig. 8C). These buried channels appear to have linked the modern Poverty Canyon system with the part of the Hikurangi Channel that extends eastwards across the plateau (Fig. 8A). Low sediment waves, with wavelengths of 2–5 km and heights of several tens of metres (Fig. 8C), extend north-eastwards towards the 15-km-wide gap between the deformation front and the Gisborne Knolls (Fig. 8A), suggesting that any modern outflow from the Poverty Canyon, and perhaps also overflows from the Hikurangi Channel, flow between Gisborne Knolls and the margin into the northern trough.

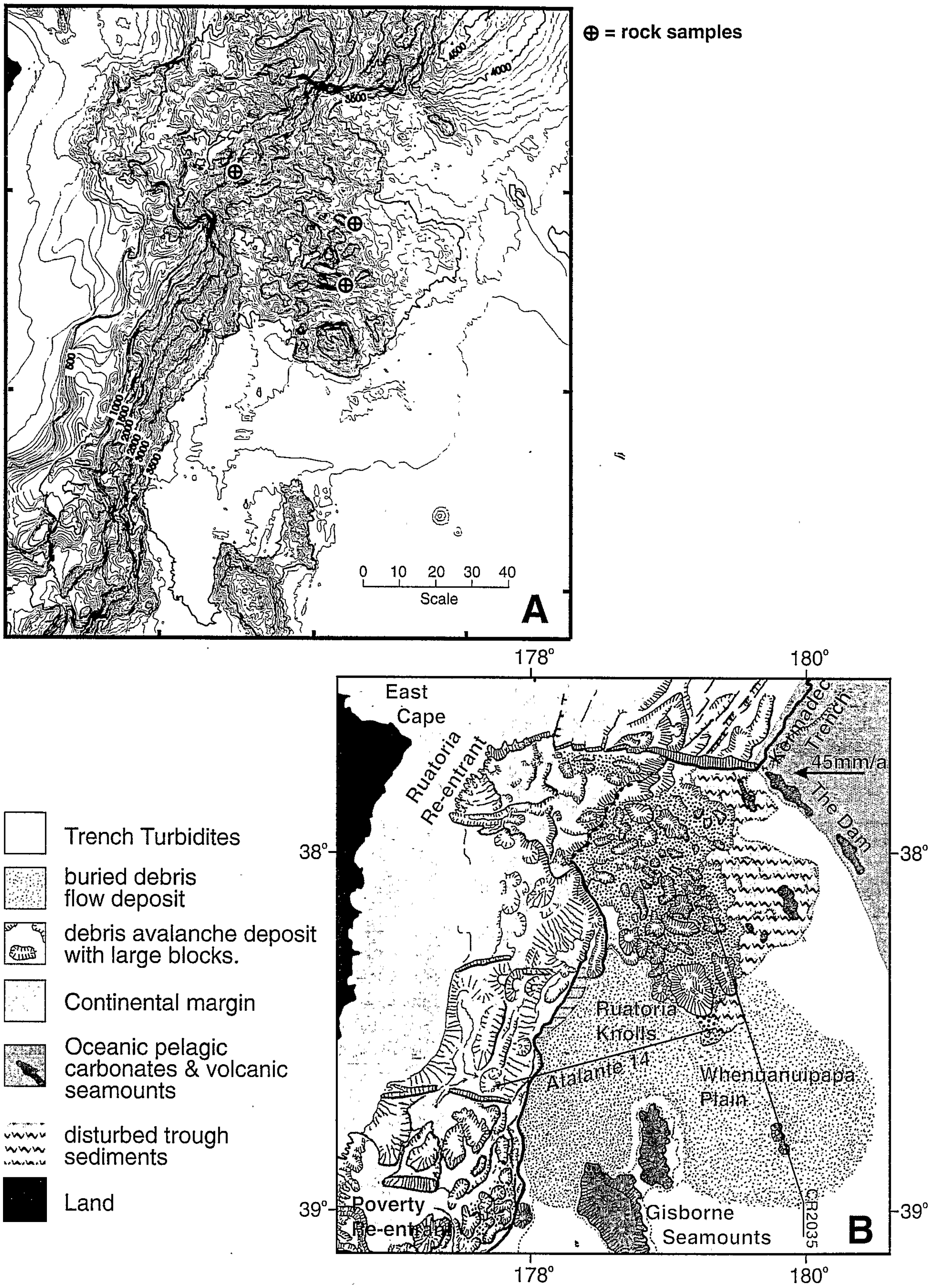
The northern trough and its avalanche and debris flow deposits

North of the gap between the Gisborne Knolls and the margin at 39°00′S, the Hikurangi Trough continues for 70 km along the toe of a steep margin (Fig. 9) that is inferred to have been eroded and over-steepened by repeated seamount impacts (Collot *et al.*, 1996). The turbidite layers of the trough are undisturbed right to the toe of the margin, which is devoid of accreted trench material (Fig. 10A). The basin depocentre contains up to about 1.0 km (1 s TWT) of mainly parallel-bedded turbidite fill that includes a conspicuous transparent (reflection-free) layer (Figs 8C and 10A,B). The absence of reflectors in this layer suggests the homogeneity of a debris flow. It thickens northwards towards an area of blocky topography and disturbed reflectors that extend out into the trough seaward of a major re-entrant in the upper margin (Figs 9A,B and 10A,B). A collapsed upper margin and a lower margin that bulges into the trough indicate large-scale slope-failure (Lewis & Pettinga, 1992). New swath bathymetry shows that slope-failure has been in the form of sliding of large blocks rather than by rotational slumping. It also shows a straight, northern scarp, subparallel to the plate convergence direction, that suggests the impact of a large subducting seamount with slope failure occurring in its wake (Collot *et al.*, 1996).

The large blocks extend, slightly obliquely, out into the trough for a distance of nearly 40 km from the toe of the slope. They infill the trough for about 40 km along its axis (Fig. 9). Seaward of a line joining the toe of the slope on either side, there are four blocks, each at least 8 km across, that rise more than 1 km above the level of the turbidite plain and another 12, each more than 4 km across, that rise at least 500 m. There are more such blocks on the lower slope. Samples from two blocks (Fig. 9A) that are about 15 km into the trough are of mid to upper Miocene mudstone, similar to those on the adjacent margin, and they both contain shelf or upper slope microfossils (A. Edwards, personal communication). They are unlike the limestone with pelagic microfossils dredged from volcanic knolls on the subducting plate (Lewis, 1985). The largest block, one of the Ruatoria Knolls, is also the most seaward. It is 17 km long by 13 km wide and rises 1200 m above the trough floor, with seismic profiles suggesting that it has landward-dipping bedding (Lewis & Pettinga, 1992) and a basal plane 400 m below the level of the adjacent plain (Fig. 10B). This mass of enormous slide blocks at the northern end of the Hikurangi Trough is considered to represent a giant debris avalanche deposit, analogous to those reported from oceanic volcanoes (Jacob, 1995; Masson, 1996).

Beneath the 100-km-long by 100-km-wide

Fig. 8. Form and structure of the Mahia depocentre and adjacent areas. A: Diagrammatic representation of morphology and backscatter showing channel and overbank bedforms, including wave-train at tangent from southern bend and another towards gap between Gisborne Knolls and northern margin, and strong backscatter from Poverty Canyon and canyon-head gullies of Poverty Re-entrant. Broken lines with arrows show buried channels between Poverty Canyon and Hikurangi Channel. B: Tracing of seismic profile across the southern Mahia depocentre showing turbidite-fill thickening towards the deformation front and onlapping onto old basin sediments and pelagic carbonate drape to the east. Vertical scale in seconds (profile Mobil 73-23). C: Tracing of seismic profile along northern Mahia depocentre showing buried, deeply incised channels crossing the trough from west to east and buried debris-flow extending south to zigzag line on map A (profile Atalante 21).


⊕ = rock samples
4500
4000
3500
800
1000
2000
3000
0 10 20 30 40
Scale
A
178°
180°
East
Cape
Ruatoria
Re-entrant
Kermadec
Trench
45mm/a
The Dam
38°
39°
Ruatoria
Knolls
Atalante 14
Whenuanuipapa
Plain
Gisborne
Seamounts
Poverty
Re-entrant
CR2035
B
Trench Turbidites
buried debris
flow deposit
debris avalanche deposit
with large blocks.
Continental margin
Oceanic pelagic
carbonates & volcanic
seamounts
disturbed trough
sediments
Land

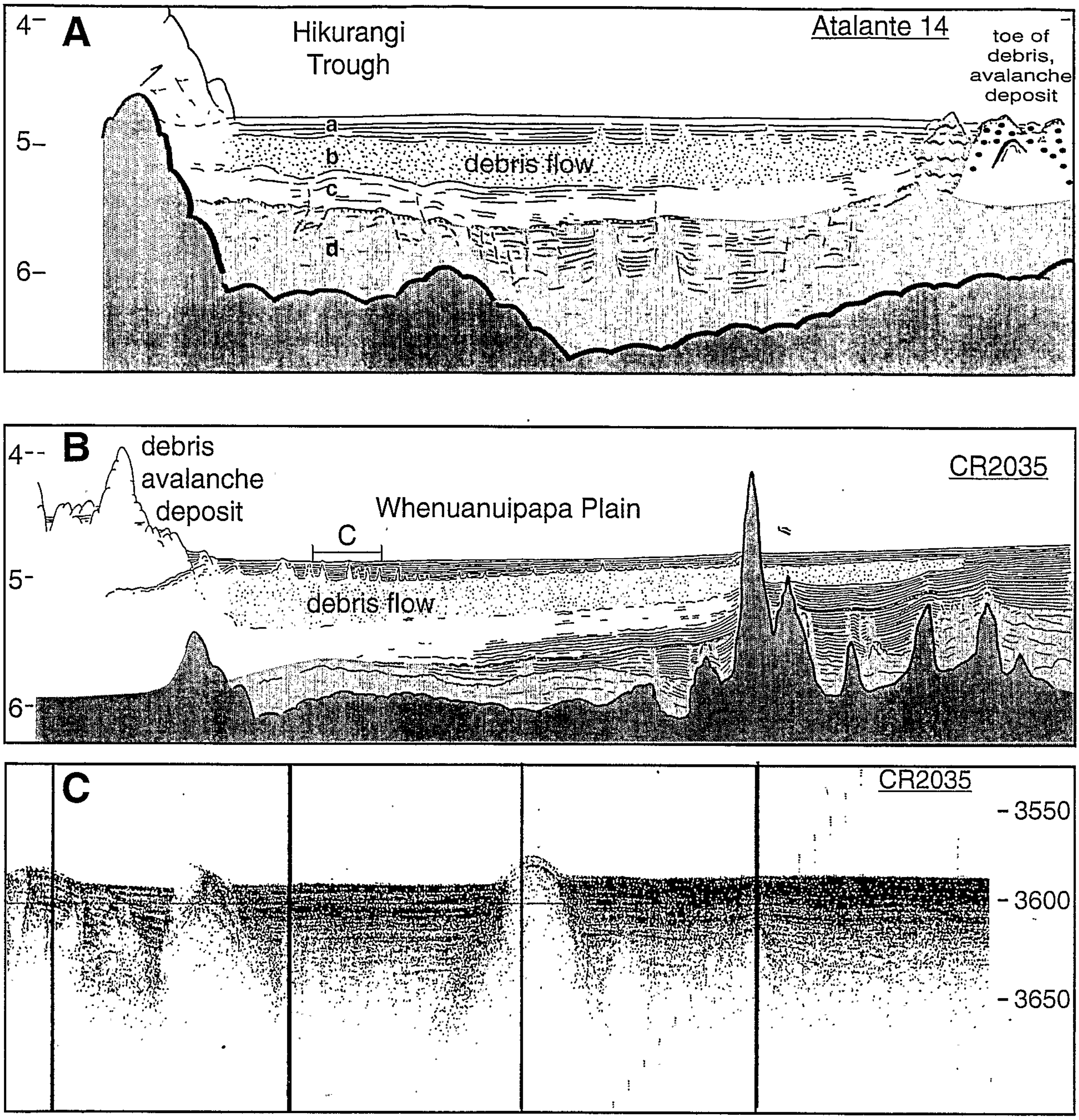


Fig. 10. Tracings of seismic profiles from the northern trough; legend as in Fig. 9. A: Profile Atalante 14 from toe of over-steepened margin (position on Fig. 9B) showing (a) sheet turbidites with small, de-watering diapirs, (b) reflection-free, debris flow deposit, (c) older turbidites and (d) pelagic drape. B: Profile of debris flow deposit beneath Whenuanuipapa Plain showing thinning of deposit away from debris avalanche deposit (position on Fig. 9B). C: High-resolution, 3.5-kHz profile of part of Profile 10B, showing diapiric 'mud volcanoes', originating from de-watering debris flow.

Whenuanuipapa Plain situated seaward of the blocky avalanche deposit, the reflection-free wedge, inferred to be a debris flow deposit, infills the trench (Fig. 10A,B). The debris flow deposit is buried by 80–180 m (100–200 ms) of turbidites that are extensively intruded by small diapiric structures from the underlying debris flow

Fig. 9. Maps of the northern end of Hikurangi Trough showing margin collapse deposits and junction with southern Kermadec Trench. A: Bathymetry with contours at 50-m intervals showing positions of Miocene mudstone samples. B: Interpretation based on bathymetry and seismic profiles (see Fig. 10) showing extent of avalanche deposit with largest blocks 40 km from toe of slope, the extent of the debris flow deposit beneath Whenuanuipapa Plain, and turbidite-fill ending at a NW/SE-trending line of seamounts that forms the present dam between trench-fill basin and sediment-starved Kermadec Trench. The blocky margin on the lower slope below Poverty Re-entrant may be a re-accreted avalanche deposit. Position of seismic profiles in Fig. 10 are shown.

(Fig. 10A–C). The diapirs are most numerous where the underlying debris flow is thickest (Fig. 10B,C). The debris flow deposit itself is about 250–300 m (300–350 ms) thick close to the debris avalanche and thins away from it to about 65 m (80 ms) thick, where it ends abruptly 70 km to the south-east (Figs 9B and 10B). It overlies a lower layer of parallel-bedded sediments that are 170–350 m thick and dip towards the margin, which in turn covers faulted 'pelagic drape' up to about 1 km (1 s) thick infilling hollows in the underlying irregular basement (Fig. 10A,B). Except for the seismically transparent, debris-flow wedge, the sequence correlates with a similar sequence in the central basin of the Hikurangi Plateau that is crossed by the eastward-trending Hikurangi Channel (Lewis, 1994; Wood & Davy, 1994). The top turbidite of the Whenuanuipapa Plain equates, at least partly, with inferred Quaternary, channel-overbank turbidites of the plateau, the lower parallel-bedded sediments partly with inferred Pliocene sheet turbidites, and the deeper 'pelagic drape' with a slowly deposited, carbonate-rich Miocene and Palaeocene layer that infills depressions in the (inferred Mesozoic) thickened oceanic basement.

Some of the diapirs penetrating the upper turbidites of the Whenuanuipapa Plain break through as mud volcanoes at the seabed (Fig. 10C), indicating prolonged and still on-going dewatering of the underlying, thick, debris flow. The limited number of reflectors beneath the debris flow suggest some disruption of underlying water-rich turbidites by rapid loading when the debris flow was deposited. Available profiles indicate that the debris flow infills the trough depocentre south of the avalanche deposit and extends beneath the Whenuanuipapa Plain for up to 100 km from the deformation front (Fig. 9B). It has a total area of about 4000 km^2. A debris flow deposit that is up to 250–300 m thick and extends out into the trough for 100 km must have been a significant, if temporary, barrier to turbidity currents from the south.

The modern seamount chain dam

The Whenuanuipapa Plain, which represents the northern limit of the Hikurangi Trough turbidite plain, ends at a NW/SE-trending chain of low seamounts (Fig. 9A,B). This seamount chain intersects the trench at 37°45′S, and its subducted parts are discernible beneath the inner trench wall (Collot *et al.*, 1996). To the north, the southern Kermadec Trench is characterized by greatly increased depths, by *en echelon* enclosed basins, by a V-shaped profile, and by the absence of turbidites, although there is still up to 1 km of pelagic-drape sediments (Collot *et al.*, 1996). The chain of low seamounts is currently the dam that separates the flat plains and turbidite basins of the Hikurangi Trough from the enclosed, sediment-starved basins of the southern Kermadec Trench.

DISCUSSION OF SEDIMENT/STRUCTURE INTERACTIONS IN A DAMMED TRENCH

Sedimentary inputs and tectonic outputs

A sediment-filled trench is a special type of sedimentary basin, one that is being continuously consumed by tectonic processes along one side. Hence, basin characteristics depend not only on the nature and rate of sediment supply and loss but also on the rate of sediment loss by subduction processes. The Hikurangi Trough is a sedimentary basin because inputs have exceeded outputs, whereas in the Kermadec Trench they have not.

At a continent-flanked trench, such as the Hikurangi Trough, sediment input is likely to be high, sourced from uplifted forearc strata onshore. Where the trench ends at a transform that includes a component of continental collision and uplift, as it does at the Hikurangi Trough, then sediment supply to the apex of the trench is likely to be particularly large.

Outputs from shallower parts of the trench can also be voluminous. In the absence of significant barriers, sediments may be transported along the trench axis to its deepest part, perhaps accumulating in small depocentres along the way. These small depocentres are likely to be consumed beneath the margin, so that a wide turbidite plan never develops. However, at the Hikurangi Trough a predominantly apical sediment supply has been contained within a restricted length of trench. Tectonic outputs by accretion and subduction increase with both increasing rates of convergence and decreasing obliquity of convergence (Underwood & Bachman, 1982; Von Huene & Scholl, 1991); decreasing obliquity implies an increasing orthogonal component of convergence. The Hikurangi Trough's rate of convergence increases, and its obliquity of convergence decreases, towards the north (De Mets *et al.*, 1990, 1994). Consequently, tectonic output by subduction processes are likely to increase in that direction and the rate of sediment supply decreases in the same direction. The net effect is that the Hikurangi Trough is a sedimentary basin with turbidite fill that thins from about 5 km thick in the south to less than 1 km in the north.

The degree to which the subducting plate is 'roughened' by tensional grabens or seamounts influences the mechanism and rate of loss by tectonic processes. A smooth basement, with few seamounts, such as occurs in the southern Hikurangi Trough (Wood & Davy, 1994), encourages offscraping of trench-fill into an accretionary prism, whereas a 'rough' basement topography with many seamounts, like the northern trough, protects sediment in depressions and enables more of it to be subducted beneath the margin (Malavieille *et al.*, 1991; Lallemand *et al.*, 1994; Mountney & Westbrook, 1996). It is estimated that of the order of 60 000 km^3 of turbidite fill remains in the Hikurangi Trough, with in excess of 20 000 km^3 of Quaternary trough turbidites preserved in the accretionary prism on the adjacent margin. The amount of trench-fill subducted beneath the margin is unknown, although in the central trough very little trench-fill is subducted, the decollement between offscraped and subducted sediment being close to the boundary between turbidite fill and the underlying pelagic drape that blankets the plateau.

Kaikoura Canyon supply to the trough

Because it is the only canyon that projects close enough to the coast to intercept a major longshore sediment transport system (Lewis *et al.*, 1998), the Kaikoura Canyon is inferred to have been the main conduit for turbidites to the Hikurangi Trough during the late Holocene (Carter *et al.*, 1982; Lewis, 1994). Longshore transport into the canyon head is estimated at be currently of the order of 1.5 km^3 kyr^{-1} (Lewis & Barnes, 1998). The rate of deposition in the southern Hikurangi Trough has been about 0.65 m kyr^{-1} during the last 0.5 Myr (Barnes & Mercier de Lépinay, 1997), and the average over the whole trough is estimated to have been about 0.4 m kyr^{-1}. The area of the trough, including the area accreted to the over-riding plate during the last 0.5 Myr (Barnes & Mercier de Lépinay, 1997), is about 37 000 km^3 and, accepting the average sedimentation rate of 0.4 m kyr^{-1}, the volume added to the trough during the late Quaternary has been about 15 km^3 kyr^{-1}. This estimate includes nonturbidite deposition, including tephra from andesite volcanoes to windward, and fine-grained sediment from rivers and resuspended from the shelf. It does not include turbidite deposition on the outer part of the Hikurangi Plateau or in the Hikurangi Channel's distal fan-drift. Nevertheless, the average rate of addition to the Hikurangi Trough is probably an order of magnitude greater than the Holocene input from the Kaikoura Canyon. Clearly, deposition was very much greater during glacial ages when many other canyons supplied sediment directly to the shelf edge and when large amounts of bedload was not trapped by glacial lakes (Carter & Carter, 1990).

The canyon-head fill consists of longshore-migrating fine sand and mud, with some locally derived gravels (Lewis *et al.*, 1998). The frequency of major collapse of the metastable pile can be gauged from the periodicity of the resultant deposits. Turbidites in the main trough, where deposition can be dated from tephra layers, suggests that major turbidity currents occur there once every few centuries (Lewis, 1994). This invites a correlation with major earthquakes, because movements of the major plate boundary faults near Kaikoura, which are capable of producing seismic shaking with a modified Mercalli (MM) intensity of more than VIII, occur at intervals of 90–400 years (van Dissen, 1991). Correlations between earthquakes, slope failure and turbidity currents are well documented and it is inferred that severe shaking of metastable sediment increases pore pressure and sensitivity to the point of failure (Houtz & Wellman, 1962; Morgenstern, 1967; Krause *et al.*, 1970; Hughes Clarke, 1990; Garcia & Hull, 1994; Garfield *et al.*, 1994).

At Kaikoura Canyon, major inputs of other types are less likely. Direct hyperpycnal inflow of turbid fluviatile water (Droz *et al.*, 1996) is impossible as there are no significant rivers near the canyon head. However, storm effects could be significant in a bay exposed to the full force of southerly storms. Storm-initiated flows from breaking internal waves and storm-generated edge-waves can reportedly resuspend large quantities of sand and generate rip-like flows (Inman *et al.*, 1976), and these could theoretically 'ignite' to reach catastrophic equilibrium like an earthquake-triggered event (Parker, 1983). However, in a zone of severe earthquakes, such events are unlikely to rival the near instantaneous failure of large volumes of sediment from ground shaking, but they may be responsible for distributing fine shelf sediments into proximal parts of the canyon where they can be incorporated into later large events.

Physical and theoretical modelling demonstrate that an initial slide in a steep canyon head rapidly metamorphoses into a turbidity current, perhaps via a debris flow stage (Mulder *et al.*, 1997), by entraining water via tunnels in the base of the initial flow and increasing the volume in motion by 2–6 times (Allen, 1971). Even as the flow becomes a turbidity current, gravel can still move by grain-flow or debris-flow processes at the base of the flow (Lowe, 1982) and the low sinuosity and steep gradient of the Kaikoura Canyon is typical of conduits with high bedload transport (Clark *et al.*, 1992; Clark & Pickering, 1996). Concurrently, the more fluid, upper flow may outpace the gravelly base (Masson, 1994), and may become supercritical, resulting in additional mixing at the upper interface. Given an erodible substrate of soft, hemipelagic and storm deposits, each major turbidity current may rapidly attain a state of self-perpetuating autosuspension (Bagnold, 1962), or even 'ignite' into a self-accelerating, ignitive autosuspension current, which erodes more than it deposits and increasing its power until a state of 'catastrophic equilibrium' is reached (Pantin, 1979, 1983; Parker *et al.*, 1986). In the Kaikoura Canyon and head of the Hikurangi Trough, exchange of sediment with the seabed probably takes place in two stages; most erosion of fine, soft sediment takes place at the head of the flow while coarse sediment is deposited from the body and tail of the flow (Simpson, 1987). The result is an ongoing exchange of sediment with the seabed, with a fluid, autosuspension turbidity current that overflows the Hikurangi Channel for enormous distances and ultimately bears little resemblance in form and composition to the flow that began in the canyon head.

In the Kaikoura Canyon, it apparently takes several centuries for sufficient metastable sediment to accumulate in the canyon head to generate a flow that will 'ignite' into a long-distance turbidity current. Minor flows may never 'ignite', but deposit their entrained sediment as 'fuel' for the next big one. Theory predicts that, in any particular morphological setting, catastrophic equilibrium flows will naturally attain a similar size, concentration and velocity, producing the uniformly sized turbidites characteristic of many modern and fossil basins (Pantin, 1983), including the Hikurangi Trough (Lewis, 1985).

Other canyons and 'headless' gullies at high and low sea-level

Although the Kaikoura Canyon is inferred to be currently the most important conduit for sediment to the trough, it was probably a relatively minor feature during low-stands of sea-level. Then, the Pegasus Canyon system intercepted enormous amounts of sand moving longshore from rivers far to the south (Herzer, 1979), little sediment then being trapped in shelf sediment prisms or in supraglacial lakes (Carter & Carter, 1990). At the same time, the extensive Cook Strait Canyon system was a major supplier of sediment from rivers that now flow into the western straits (Lewis *et al.*, 1994), with sediment from Cook Strait reaching the fan-drift at the far end of the Hikurangi Channel (Carter & Mitchell, 1987). Like most of the world's canyons, the Pegasus and Cook Strait canyons, which carried large quantities of traction load during low-stands, carry little during high stands (Griggs *et al.*, 1970; Carter & Mitchell, 1987; Thornberg & Kulm, 1987).

The small canyons that form at the confluence of semiregularly spaced gullies, some on midslope ridges, owe little to coastal sources. The gullies resemble those attributed to 'spring sapping' at fluid seeps (Orange & Breen, 1992; Orange *et al.*, 1994), the regular spacing of the seeps being caused by a self-regulating feedback of the excess pore pressure gradient in the underlying rocks. Seeps of methane-rich fluids, with vent faunas and carbonate crusts or chimneys, occur widely along the Hikurangi Margin, generally associated with compressional deformation of Neogene sediments (Lewis & Marshall, 1996). Seep gullies are typically aligned along faults or outcrops of porous sediment (Fujioka & Taira, 1989; Henry *et al.*, 1989; Lallemand *et al.*, 1992; Le Pichon *et al.*, 1992), and those of the southern Hikurangi Margin are aligned along fault-controlled scarps, where fluids are expelled from compressed rocks to landward (Lewis *et al.*, 1986; Barnes *et al.*, 1998). The gullies, which begin as shear strength is reduced close to a seep site, develop by headward sapping above the seep and by erosion below it. This may generate enough rock-debris to produce the strong backscatter seen in the small canyon axes. Although gullies that incise the shelf break may be enhanced by funnelling of nearshore sediments during lowstands of sea-level, and despite the number of gullies along the margin, their input to the Hikurangi Trough is likely to be small.

Canyons that incise the upper slope landward of the accretionary prism supply little to the trough because their load is trapped in the baffle of slope basins (Lewis, 1980). Further north, the large Poverty Canyon system reaches the trough without any such baffle, the low ridge at the outer edge of the small triangular ponded basin at its toe being an insignificant barrier to high-velocity turbidity currents (Underwood, 1986; Garcia & Hull, 1994). With the virtual absence of a left-bank levée on the Hikurangi Channel opposite the Poverty Canyons, flows might be funnelled directly into the segment of the channel that crosses the plateau to the east. However, sediment waves in the trough to the north-east of the Poverty Canyon toe suggest that at least some flows from Poverty Canyon may turn left into the northern end of the Hikurangi Trough, perhaps under the influence of either Coriolis forces or a shallow branch of the Deep Western Boundary Current that crosses the Hikurangi Plateau (McCave & Carter, 1997). Although the head of the Poverty Canyon is sufficiently deep to have only been a major conduit for detrital sediments during periods of glacially lowered sea-level, regularly spaced, strongly reflective gullies along the straight shelf edge imply input from seeps, which are known to occur both to seaward and to landward (Lewis & Marshall, 1996). Thus, although Poverty Canyon may have been the only significant source of turbidity currents north of Cook Strait its main input was during glacial ages, with minor on-going input from seep-gullies.

The absence of fans

It has been observed that well-developed, submarine fans form in trenches at the bottom of steep canyons, although they form recognizable features only if their rate of development exceeds the rate at which they are being destroyed by subduction or accretion (Thornberg & Kulm, 1987; Thornburg *et al.*, 1990). They are inevitably transitory features. It has also been observed that trenches with well-developed axial channels do not have well-developed submarine fans (Underwood & Bachman, 1982). The Hikurangi Trough, with its well-developed channel and absence of fans, conforms to this observation. Nowhere along the Hikurangi Trough are there large, canyon-toe, distributary fans, similar to those of some other modern and ancient turbidite-filled trenches (Mutti & Normark, 1987); the faint, fan-shaped backscatter patterns on the flat trough floor beyond some small canyons (Fig. 4A) have little or no bathymetric expression. Another analogue may be the sediments that are confined in the head of the Hikurangi Trough between the flank of the Chatham Rise and margin of north-eastern South Island. They may represent a structurally confined and deforming trench-apex fan, similar to the Boso Fan at the head of the Sagami Trough of Japan (Soh *et al.*, 1990). Like fans elsewhere, the trough-head sediments are at the toe of canyons where coarse sediment is deposited at a reduction in slope, perhaps because of a hydraulic jump in large flows. The hydraulic jump entrains water, reduces effective density, reduces velocity (Rothwell *et al.*, 1992), encourages deposition of coarse sediment, increases flow thickness, increases turbulence, increases the amount of fine sediment that can remain in suspension (Kenyon *et al.*, 1995) and may assist formation of turbidity currents capable of long-distance, self-perpetuating flow (Komar, 1971; Pantin, 1979).

We have no reliable estimates of deposition rate in the head of the trough, but further north, Holocene rates of deposition, based on the depth to dated rhyolitic tephras, are about 0.15 m kyr^{-1}, with one turbidite every 1000–1500 years (Ninkovich, 1968; Lewis & Kohn, 1973); rates increase to 0.3 m kyr^{-1}, with one turbidite every 300–400 years, on a right-bank levée (Lewis, 1985, 1994). However, average rates of overbank deposition in the southern trough, averaged over several glacial–interglacial cycles, are more than 0.65 m kyr^{-1} (Barnes & Mercier de Lepinay, 1998). Such rates of deposition are moderately rapid for a deep-water basin (Rothwell *et al.*, 1992), although slow compared with large fans (Damuth & Flood, 1984; Feeley *et al.*, 1990).

Bedforms and flow in an axial channel

The conspicuous channel-axis bedforms of the Hikurangi Trough are indicative of the flows with which they are in equilibrium, although the relationship is not yet well quantified. Transverse sediment waves are probably formed by large eddies, which in turn may be related to thickness of the flow (Yalin, 1972; Zeng *et al.*, 1991). The ratio between bedform wavelength and flow thickness has been inferred to range from 7 : 1 to 1 : 1 (Allen, 1984; Zeng *et al.*, 1991). Therefore, the 3.5-km wavelength bedforms that occur in the channel axis before the confluence with the Cook Strait Canyon (Fig. 4A) were formed by flows at least 500 m thick. Those beyond the Cook Strait confluence, which have a wavelength of 2.5 km, were formed by flows at least 350 m thick. Small superimposed bedforms with a wavelength of 600 m may relate to the tails of large flows or to small flows that were at least 90 m thick. Transverse bedforms on the insides of meanders, although comparable with those on fluviatile point-bars (Miall, 1989), have a wavelength indicate of flow thicknesses greater than the channel depth.

One of the enigmas of long-distance autosuspension currents in the Hikurangi Trough is that they must erode soft sediment from a channel that is aggrading. Entraining new, fine sediment into the flow, which is critical for the continuation of the flow, is an important mechanism of sediment transport (Piper & Stow, 1991). Entrainment is quantified for the 1929 Grand Banks turbidity current, which entrained 50–100 km^3 of sediment on the lower slope (Hughes Clarke *et al.*, 1990), and is well documented for channels in supraglacial lakes that are both cut and filled during passage of a turbidity current (Eyles, 1987). Mainly deep-water, rather than shallow-water, foraminifers occur in distal turbidites of the Hikurangi system (Fenner *et al.*, 1992), demonstrating that trough mud is entrained and carried to more distal areas along with continually reducing proportions of the original shelf sediment.

In the Hikurangi Channel, where does the mud that is entrained into long-distance flows come from? Little comes from the erosion of channel floor and walls because the channel system aggrades and, since there is no evidence of rapid meander migration, any erosion of the outsides of bends is compensated by deposition on point bars. The lack of extensive meander migration, evinced by abandoned meander loops and systems, indicates that the Hikurangi Channel is broadly in equilibrium with the prevailing flow conditions (Shor *et al.*, 1990). There is some from collapse of channel walls (Fig. 4) but, although this makes a significant contribution to some channel systems (Masson, 1994; Masson *et al.*, 1995), it is here localized and does not provide an ongoing supply along the channel. Alternatively, much of the sediment required to 'fuel' the larger flows may be input to the channel by dilute, turbid flows that never 'ignited'. Such dilute flows may begin during storms on the narrow shelves of the region, flow down the margin and deposit thin watery clay-rich layers in bathymetric lows (Lewis, 1973; Lewis & Kohn, 1973). Such layers may build up over centuries to form the 'hemipelagic' mud, which forms 30–50% of the mud in some cores from the turbidite plain (Lewis, 1985), although such deposits may more appropriately be termed 'hemiturbidite' (Howe, 1996). The lowest place for such watery, muddy 'hemiturbidite' layers (Howe, 1996) to accumulate is the channel itself. Over time, a significant layer of fine sediment may accumulate in the channel ready for the next autosuspension flow, which entrains the mud and deposits coarser sediment in its wake.

There is debatable evidence of a significant layer of fine sediment in the channel at the present time. Most of the channel in the Hikurangi Trough has lower backscatter than adjacent overbank areas (Fig. 4C,D). Generally, though not universally, lower reflectivity is indicative of finer grained sediment at the seabed. This indication of fine-grained sediment in the axis of the channel appears to contradict the evidence from cores, which show that graded silty sand and sandy silt turbidites are thicker and more abundant in the channel axis than on the surrounding plain (Lewis, 1985). However, in one core from the southern trough, the top turbidite is buried by over 0.4 m of soft 'hemiturbidite' mud (P. M. Barnes, personal communication), which may have been lost in the coring process from other archived cores and perhaps partly even from this core. Thus, the low backscatter from the channel axis may reflect a surface layer of soft, fine sediment. Despite the generally positive correlation between backscatter intensity and grain size, constructive and destructive interference in well-layered sediments can produce anomalous results (Gardner *et al.*, 1991) as can biological activity, which increases surface roughness and backscatter from fine sediments in overbank areas (Kenyon, 1992). Hence, the swath imagery of a weakly reflective channel suggests, but does prove, the presence of a significant layer of soft mud sitting within the channel that is waiting to be entrained in the next catastrophic flow.

Channel sinuosity is likewise indicative of flow, with high sinuosity being incompatible with very fast-moving turbidity currents (Klaucke *et al.*, 1997). The sinuosity/gradient profile of the Hikurangi Channel, which has a maximum sinuosity of 1.5 and a channel slope decreasing from 1 : 400 to 1 : 800 in the southern Hikurangi Trough, is in the midrange for canyon/channel systems and is characteristic of silt-dominated flows; the Cascadia system, which is also at a convergent margin, has a similar profile (Clark *et al.*, 1992).

Flows out of and into the channel

The form of the trough's channel levées reflects the muddy nature of the upper part of equilibrium flows along the channel. Since the left bank levée is consistently higher than the right, even at left-hand bends (Lewis, 1994), the implication is that southern hemisphere Coriolis effects have consistently exceeded centrifugal effects in the elevated heads of thick turbidity flows (Carter & Carter, 1988). Such consistent cross-flow asymmetry is achieved in moderately slow and dilute flows (Komar, 1969; Piper & Savoye, 1993; Klaucke *et al.*, 1997). Nevertheless, there is strong evidence of centrifugal outflow of some flows.

The 5–15-m-high, 2–3-km-wavelength mud-waves that occur tangentially to the outsides of bends on both banks (Figs 4D, 6B) is evidence of strong, centrifugal outflows. Centrifugal effects probably occur in the heads of larger, denser, faster turbidity currents (Griggs & Kulm, 1970). A detailed consideration of turbidity current flow parameters is beyond the scope of this paper. There are many variables, and assumptions must be made about many of them. Analyses of some turbidite systems have been made on the basis of grain size of the deposits (Van Tassel, 1981; Komar, 1985) but this method has been questioned (Hiscott, 1994). For others, interpretations of flow characteristics have been made from channel dimensions (Komar, 1969). However, a generalized formula for a 500-m-thick autosuspension current, where gravity is balanced against friction, suggests that velocities at the head of a moderate-sized flow, may reach about 30 m s^{-1} in the lower Kaikoura Canyon, 12 m s^{-1} in the restricted upper trough, 6 m s^{-1} in the southern trough near Cook Strait, and 4 m s^{-1} off Mahia, and thereafter increasing to 6 m s^{-1} as the axial slope increases after the channel leaves the trough (H. M. Pantin, personal communication).

The weakly backscattering mudwaves that occur along, and parallel to, the levée crest on the higher left bank levée only probably reflect outflow that was strongly influenced by Coriolis forces, for instance, the thick, dilute and slowly moving tail-plumes of major turbidity currents or some dilute, nonautosuspension flows that occur between the major flows. Similar levée-parallel mudwaves occur beside the river-fed Zaire Channel (Droz *et al.*, 1996).

Turbidity currents may considerably overlap a channel but still follow its course. 'Intermediate' flow are perhaps twice as deep as the channel and exceptionally large ones are 10 times the channel depth (Clark & Pickering, 1996). In the Hikurangi Trough, where the channel is 150–200 m below the right-bank levée and 200–280 m below the left bank levée, an 'intermediate' flow might equate to the > 500-m-thick flow inferred to form the channel axis bedforms. For such a flow, loss of the fine, dilute upper part of each turbidity current to overbank areas, a phenomenon known as flow stripping (Piper & Normark, 1983), could continue for hundreds of kilometres, producing the thin silt turbidites that have been correlated elsewhere over vast areas (Hess, 1995; Klaucke *et al.*, 1998). Continuous flow stripping requires that the channel be in dynamic equilibrium with the turbidity currents that are typical of this system. The overbank flow of the channel may form secondary, sheet flows along the toe of the margin, producing margin-transverse bedforms (Figs 4A and 6A). In some instances, the stripped flow may rejoin the main channel flow at places with reduced levées (Rothwell *et al.*, 1992; Masson *et al.*, 1995); in some case, the channel may recapture only the coarsest part of the flow (Masson, 1994). In the Hikurangi Trough, overbank flows probably rejoin the channel at several places where the channel approaches the margin and its left bank levée is low, notably where seamounts and protothrusts fill the gap between channel and margin (Fig. 7C) and immediately before it turns east out of the trough (Fig. 8A).

Where the channel turns right out of the trough, turbidites indicate that sheet flows continue northwards along the toe of the margin and detour around the giant avalanche deposit for 170 km to the northern end the trough (Figs 2 and 8). On the central basin of the Hikurangi Plateau, sheet-flows were inferred to have travelled eastwards for more than 300 km prior to its being cut by an extension of the Hikurangi Channel (Fig. 1; Lewis, 1994). Elsewhere, unconfined, overbank, sheet-flows, which deposit their loads slowly as they slow down, are even more widely dispersed (Pickering *et al.*, 1989; Clark & Pickering, 1996), some having travelled for over 1000 km and over significant bathymetric highs (Garcia & Hull, 1994).

The Ruatoria avalanche and debris flow deposits

Although autosuspension currents can propagate for enormous distances, at least part of the reason they have never reached the southern Kermadec Trench may be damming by avalanche and debris flow deposits. At the northern end of the Hikurangi Trough, a huge, blocky avalanche deposit and associated massive debris flow deposit have infilled the northern end of the basin. These deposits were large enough to act as an 'earth dam', which, for a time during the Pleistocene, ponded turbidity currents coming from the south and prevented their passage to the southern Kermadec Trench. Structural details of the failure, its possible causes, and its implications for margin evolution will be presented elsewhere; here, we concentrate on those aspects that are directly related to the evolution of the trough.

The margin collapse into the northern end of the trough added a huge amount of fill into the basin. The avalanche deposit, which extends 40 km out into the trough, covers an area within the trough of about 1500 km^2 and constitutes an estimated 1500 km^3 of basin fill. This is only part of the total avalanche deposit, which also covers an extensive area of the lower slope in blocky topography. Its original extent within the basin may have been larger because the subducting plate is carrying it back towards the margin at the convergence rate of 45 km Myr^{-1} (Fig. 1). In addition to the blocky avalanche deposit, the debris flow deposit, which now extends 100 km out into the basin, covers an area of 4000 km^2 and constitutes an estimated 800 km^3 of basin-fill. Although the avalanche deposit and debris flow deposit together constitute a massive slope failure, they are not the world's largest. Margin collapses off South Africa are nearly an order of magnitude larger (Dingle, 1977, 1980), but may not be single features. Both the main Storegga Slide off Norway (Bugge *et al.*, 1987, 1988; Jansen *et al.*, 1987) and the Nuuanu debris avalanche of the Hawaiian Islands are about twice as big (Jacobs, 1995; Moore *et al.*, 1989). Even the 13 × 17-km slide block that has moved at least 40 km out into the basin is not the world's largest slide-block, being only about one-third of the volume of a volcanic block, Tuscaloosa Seamount, that collapsed from the flank of the Hawaiian Islands (Jacobs, 1995).

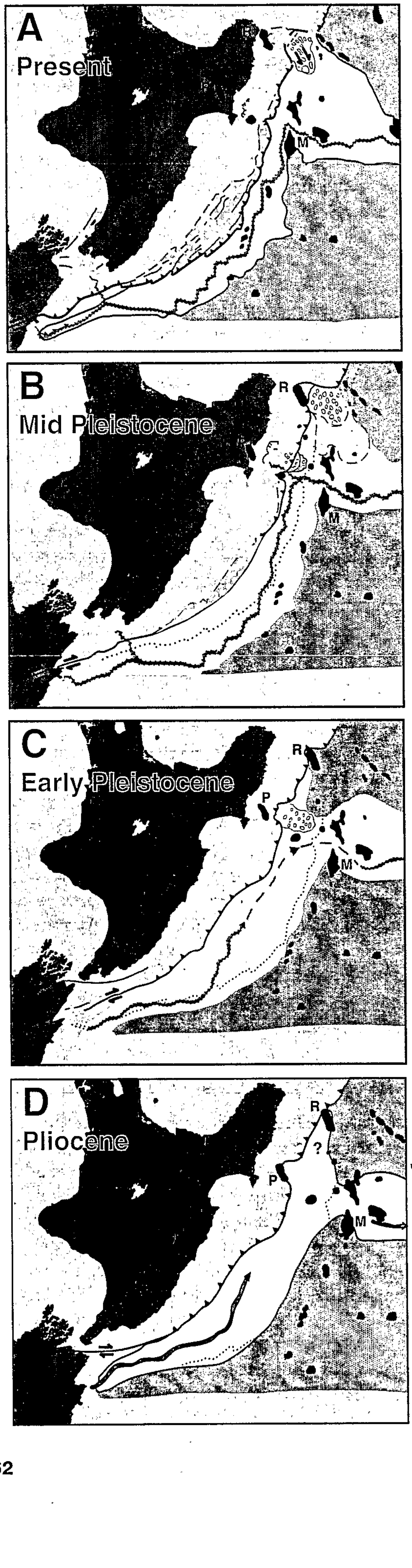
Together, avalanche and debris flow deposits infilled the trench axis and are inferred to have acted as a major barrier to the passage of turbidity currents. The avalanche deposit raised the level of the trough axis by a minimum of 500 m and the debris flow deposit raised it by over 250 m at its inner edge and by 65 m at 100 km out into the basin. It must have taken a considerable time for turbidites to overtop the debris flow and then bury it with 80–180 m of sediment. There are no reliable estimates of long-term sedimentation rates in this area but in the southern part of the trough, the rate of accumulation for the last few glacial–interglacial cycles, based on seismic stratigraphy, is about 0.65 m kyr^{-1} (Barnes *et al.*, 1998) in an area where Holocene rates are 0.15 m kyr^{-1}. Rates of Holocene (and latest Pleistocene) turbidite/hemiturbidite deposition on the central Hikurangi Plateau are inferred to be only about half that in the southern trough (Fenner *et al.*, 1992). If average, long-term rates of deposition in the northern trough are less than in the southern trough and more than the central plateau, due to greater ash input, then the long-term average may be about 0.5 m kyr^{-1}. If this value is accepted as a working approximation, then the 150–350 m of sediment that it took to overtop the debris flow and then bury it required of the order of 300–700 kyr. Thus the debris flow and perhaps also the avalanche deposit are tentatively inferred to be mid Pleistocene in age (Fig. 11A,B). Elsewhere, the largest slide blocks in debris avalanches are concentrated in the middle of a deposit (Jacobs, 1995; Masson, 1996), so that the presence of the largest slide blocks in the Hikurangi Trough at the leading edge of the avalanche deposit could imply a two-stage process whereby the large blocks slid from the margin first, and were later pushed along in front of the main avalanche. Debris avalaches are thought to be catastrophic events that move disaggregated blocks for many tens of kilometres, and up to 230 km, across an oceanic basin (Moore *et al.*, 1989), generating enormous tsunamis (Moore & Moore, 1984; Dawson *et al.*, 1988). In contrast, deep-seated, rotational slumps move relatively intact blocks for up to several tens of kilometres over a long period of time (Moore *et al.*, 1989). If this is correct, then in the northern Hikurangi Trough, an initial slow slump may have been pushed along in front of a subsequent thin, fast-moving, blocky avalanche.

From the geometry of the re-entrant's northern scarp, it has been inferred that the Ruatoria Re-entrant is primarily a seamount impact scar (Collot *et al.*, 1996) with margin collapse of collision fractured rocks in its wake (Collot & Fisher, 1989; Lallemand *et al.*, 1990; Von Huene & Lallemand, 1990). The implication for the evolution of the Hikurangi Trough is that a subducting seamount can dam a trench from the time that it first begins to imbricate trough sediments between itself and the margin until the debris that collapsed in its wake is carried back into or beneath the margin by the subducting plate. If the leading edge of the seamount or line of seamounts that formed the Ruatoria Re-entrant is ahead of the scar, then it must be 90 km or more landward of the toe of the margin in the direction of relative plate convergence. This implies that, with convergence of 45 km Myr^{-1}, the seamount first began to dam the trench at about 2 Ma, that is, in latest Pliocene or early Pleistocene times (Fig. 11C,D), and its wake effects are still limiting propagation along the trough to some extent today (Fig. 11).

The past and present dams

Turbidites are almost level with saddles in the line of low seamounts that limits the flow of distal, sheet turbidity currents into the empty Kermadec Trench (Figs 9A and 11A). The present dam is about to be overtopped by accumulating turbidites. Nevertheless, the seamounts may still act as an effective barrier because the distal turbidity currents will have lost most of their energy after negotiating 800 km of channel and a 170-km-long, circuitous route around seamounts and avalanche deposits since leaving the channel, any remaining energy being dissipated through reflection and interference at the line of seamounts (Pantin & Leeder, 1987).

The debris flow deposit, which infills the gap between



the Gisbourne Seamounts and the margin and extends 100 km out beneath the Whenuanuipapa Plain, must have been an effective barrier to the passage of turbidity currents from the time it was deposited at about 300–700 ka until it was overtopped and buried. Assuming, as suggested earlier, that the average sedimentation rate in the area is of the order of 0.5 m kyr^{-1}, then the 80–180 m of turbidites that now bury the debris flow have been deposited in the last half million years. Thus, the debris flow acted as a dam from about 0.5 Ma to about 0.25 Ma (Fig. 11A,B). If it is correct that subduction of a large seamount or NW/SE-trending chain of seamounts triggered the margin collapse, and that its leading edge began damming the trench at about 2 Ma with its trailing edge disappearing beneath the margin at about 1 Ma, then the seamount and its wake avalanche/debris flow deposits acted as an effective barrier to along-trench, sheet turbidity currents from about 2 Ma until about 0.25 Ma. Therefore, a large seamount and its wake deposits dammed the Hikurangi Trough for a period of 1.75 Myr.

Turbidity currents were being rerouted out of the trench axis and across the Hikurangi Plateau east of Mahia for perhaps several million years before the seamounts of the Ruatoria re-entrant could have acted as a dam (Lewis, 1994). What caused the diversion of flow? The 20-km-deep, V-shaped, re-entrant in the deformation front off Mahia (Fig. 8A), with its re-curved accretionary ridges, is diagnostic of a scar left by a seamount that is only just disappearing beneath the margin (Masson *et al.*, 1990; Lallemand *et al.*, 1994). If its trailing edge is about 30 km landward of the undeformed deformation front and its leading edge is 20–30 km further landward, then this seamount may have restricted flow along the trough axis from about 1.5 Ma to less than 1 Ma. Thus, it is far too young to have first caused diversion of turbidity currents out of the structural trench; in fact, its effect as a barrier overlaps with the seamount and wake deposits to the north. Significantly, the Poverty Re-entrant appears to be not a single feature, but a double feature caused by two quite separate seamount impacts (Collot *et al.*, 1996). The main Poverty Re-entrant has a 40-km-long, gullied headwall that is too large and too far north to have been

Fig. 11. Inferred evolution of the Hikurangi Trough during the last few million years. Legend as in Fig. 1. Dotted lines in B, C and D are traces of present deformation front on Pacific Plate to show area to be subducted. Note (i) turbidite basin being continuously consumed at plate boundary along its whole western edge and reforming by onlap along its eastern edge; (ii) growth of accretionary prism from basin turbidites and consequent change in alignment of trough and plate boundary; (iii) effect of subducting parts of original channel, particularly at site of buried channel segments off Mahia; (iv) progression of large seamounts and formation of wake avalanche and debris flow deposits that have together restricted flow of turbidity currents to the Kermadec Trench.

formed by the seamount that formed the V-shaped indentation in the deformation front at its southern edge (Fig. 8A). It has an almost circular, midslope basin that is characteristic of a very much older, almost healed, impact scar formed by a seamount that is now many tens of kilometres to landward of its healing re-entrant (von Huene & Lallemand, 1990; Lallemand *et al.*, 1994). In addition, it has a bulge in the lower slope, with very blocky ridges, that may represent a debris avalanche that has been accreted from the subducting plate back onto the over-riding plate (Lewis *et al.*, 1997). This almost healed scar and re-accreted avalanche deposit is clearly much older than the Ruatoria Re-entrant to the north and is inferred to have been formed by a seamount and its wake avalanche deposits that dammed the trough to the passage of turbidity currents to the north during the late Pliocene (Fig. 11C,D). It must have, at least for a time, re-routed turbidity currents out of the structural trench, eastward across the central basin of the Hikurangi Plateau.

Between the Poverty and Ruatoria re-entrants, the whole margin has been tectonically eroded and the deformation front moved landward by the impacts of many seamounts (Collot *et al.*, 1996); transverse scarps (Figs 2 and 8A) on the margin may be ghosts of seamount impact scars even older than the Poverty Re-entrant. This evidence for a succession of seamount impacts going back several million years suggests that the northern Hikurangi Trough was blocked by these seamounts and their wake deposits, more-or-less continuously, since perhaps early Pliocene times.

Wider effects of damming the trough

Damming of the trough axis to the passage of turbidity currents for several million years promoted in-filling of the structural trench to form the wide, turbidite plain, the Hikurangi Trough. An indirect consequence of thick deposits on the subducting plate was development of a wide accretionary prism adjacent to the central part of the trough (Fig. 11). Damming of turbidity currents in the trough axis also prompted overflow onto a low part of the Hikurangi Plateau to the east and infilling of a basin there (Lewis, 1994). Beginning at the end of the Miocene, sheet turbidity currents were first ponded behind a Miocene drift at the plateau edge and then eroded by a channel that cut back from the plateau edge as the drift was overtopped. By analogy with an outflow channel from a ponded basin off California (Masson *et al.*, 1995), the plateau-edge channel is inferred to have cut headward to eventually linked with the axial channel of the Hikurangi Trough to form one continuous, 1400-km-long channel between Kaikoura and the Southwest Pacific Basin. At the far end of the channel, a distal fan deposit is elongated northwards along the bottom of the Hikurangi Plateau's north-eastern scarp by the Pacific Ocean's Deep Western Boundary Current. The elongate 'fan-drift' extends to the edge of the central Kermadec Trench (Carter & McCave, 1994), which has a thin turbidite–contourite deposit in its axis (Collot *et al.*, 1996), unlike the bare southern Kermadec Trench. Thus, the trench-axis turbidity currents of the Hikurangi Trough, which subducting seamounts and their wake avalanche and debris-flow deposits have prevented from reaching the southern Kermadec Trench, eventually reach the much deeper central Kermadec Trench, but they take a very long route to get there.

CONCLUSIONS

1 The Hikurangi Trough is mainly a sediment-filled trench basin with a narrow, intracontinental, oblique collision foredeep basin at its southern apex.

2 Trench fill is mainly overbank turbidites that reach 1–5 km thickness along the western edge, where the basin is being consumed by frontal accretion, with precursor protothrusts in the south, and by subduction beneath a tectonically eroded margin in the north.

3 At its northern end, the broad, flat plain of the Hikurangi Trough changes abruptly to the narrow, V-profiled, southern kermadec trench because the northward passage of distal turbidity currents along the trough is dammed by a line of low seamounts.

4 The northern part of the trough has been dammed for several million years by a succession of subducting seamounts and by massive avalanche and debris flow deposits that collapse from the margin in the wake of the seamounts.

5 At present, the main sediment input to the Hikurangi Trough is via the Kaikoura Canyon, at its southern, intracontinental apex, as this in the only canyon that intercepts a major nearshore sediment supply during interglacial ages.

6 During glacial ages, several extensive canyon systems, which begin at the modern shelf break, funnelled sediment into the trough at a rate about an order of magnitude greater than at present.

7 There is additional minor input from headward-sapping of regularly spaced seep gullies at several fault-controlled scarps along the margin.

8 There are no clearly defined depositional fans in the Hikurangi Trough, although coarse, gravelly sediment dumped at the end of canyons into the narrow, intracontinental head of the trough may be considered partly analogous.

9 A well-developed, meandering, trench-axis channel system has bedforms indicating the passage of turbidity currents that considerably overlap the channel banks.

10 Centrifugal outflow of the heads of high-velocity turbidity currents produces trains of sediment waves aligned tangentially to the outsides of bends.

11 Coriolis effects (and perhaps deep boundary flows) on thick, dilute flows, including tails of turbidity currents, produces a left bank levée that is higher than that of the right banks, and levée-parallel sediment waves.

12 Some large turbidity currents, perhaps mainly those

triggered by rupture of large plate-boundary faults, are inferred to reach and then maintain a steady state of high-velocity, self-perpetuating autosuspension by eroding soft mud at their heads and depositing coarser, graded, turbidite layers from their body and tail. Fine sediment in the channel axis may have accumulated from dilute, muddy 'hemiturbidite' flows since the preceding autosuspension flow.

13 Basin and margin are interactive. In the southern part of the trough, where there is thick basin-fill, slow orthogonal convergence, and few seamounts on the downgoing plate, trough turbidites are frontally accreted to form a wide accretionary prism that traps sediment in slope basins and reduced lateral input to the trough. In the northern part of the trough, where there is thin basin-fill, more rapid orthogonal convergence and numerous seamounts, trough sediments are subducted and the margin is being tectonically eroded, and fractured by seamount impacts, which promotes collapse of massive avalanche and debris flow deposits into the basin.

14 The most recent margin collapse was into the northern end of the trough at about 0.5 Ma. The avalanche deposit now projects 40 km into the trough, with about 16 blocks more than 4 km wide and 500 m high added to the basin-fill, the largest being 13×17 km and nearly 2 km thick. A debris flow deposit in front of the avalanche deposit extends 100 km into the trough and thins from 250 m thick near the avalanche deposit to about 65 m thick at the seaward edge. Blanketting turbidites, up to 180 m thick, are intruded by numerous dewatering diapirs.

15 It is inferred that the thick debris flow deposit was over-topped at about 0.25 Ma and the seamount that promoted margin collapse first touched the margin at about 2 Ma. Therefore, a large seamount and its wake avalanche/debris flow deposits may block the passage of turbidity currents along the trough for about 1.5–2 Myr.

16 For about the last 5 myr, seamounts and their wake avalanche/debris flow deposits have dammed the northern Hikurangi Trough to the passage of turbidity currents into the southern Kermadec Trench, causing them to be rerouted across the central Hikurangi Plateau and into the path of the deep western boundary currents, which carries some sediment into the central kermadec trench. The southern kermadec trench has been bypassed.

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