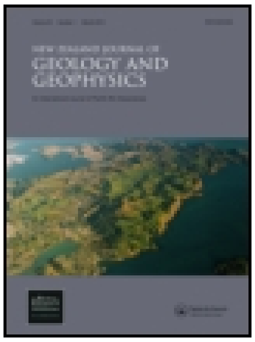

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Mud sedimentation on the continental shelf at an accretionary margin—Poverty Bay, New Zealand

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Abstract Sediments on the continental shelf, atop the accretionary prism of the eastern North Island, are dominated by mud. This situation reflects a highly erodible provenance of soft Tertiary sediments, active tectonism, meteorological extremes, and, in historical times, changing land use.

Off Poverty Bay, mud is supplied by the Waipaoa River, New Zealand's fourth largest river in terms of sediment supply. Under normal conditions, suspended sediment is dispersed as surface or hypopycnal plumes that have a net northeastward or southward dispersal along the shelf, mainly in response to the prevailing wind-driven circulation. During extreme floods with return periods of 10 years or more, fluvial suspended sediment concentrations are probably high enough to form subsurface or hyperpycnal plumes that move and disperse under gravity and shelf currents. After the 100 year Cyclone Bola event of 1988, reef communities of the inner shelf were temporarily inundated by a fluid mud layer.

Surficial sediments and 3.5 kHz seismic reflection profiles reveal that mud accumulates in a subsiding synclinal basin occupying the middle shelf. Offshelf dispersal is hindered by the growing Lachlan and Ariel Anticlines along the outer shelf. As a result, 20 km^3 of mud has been deposited since c. 18 ka, of which 8 km^3 accumulated since c. 8 ka. This late Holocene rate is nearly five times lower than the modern rate of mud supply, which equates with the marked increase in terrestrial erosion following European deforestation in the late nineteenth century.

Keywords continental shelf; mud sedimentation; accretionary margin; Poverty Bay

INTRODUCTION

The continental margin, east of the North Island of New Zealand, receives a large volume of fine terrigenous sediment, as seen by extensive mud deposits on the continental shelf (e.g., Pantin 1966; Pantin & Gibb 1968; Lewis 1973). Local rivers, such as the Waiapu to the north and the Waipaoa, have some of the highest discharges of suspended load in New Zealand, even though their catchment sizes are modest at c. 2000 km^2 (Griffiths & Glasby 1985). This influx of mud is the collective response to: (1) an abundance of highly erodible rocks in provenance areas (Trotter 1988); (2) pronounced seismicity that stems from the location of the eastern North Island within the actively deforming Hikurangi subduction margin (Lewis 1980; Reyners 1989); (3) deforestation related to European settlement from 1880 onwards (Singleton et al. 1989a, b); and (4) the passage of tropical cyclones (Sinclair 1993).

The devastating effects of these combined factors were highlighted in 1988 when the northern North Island was struck by Cyclone Bola. This 100 year event was accompanied by intense rainfall. Up to 900 mm fell in 72 h (Singleton et al. 1989a; Sinclair 1993) causing local rivers to discharge several times their mean annual sediment loads. As a result, the continental shelf off Poverty Bay (Fig. 1) was inundated by mud. Although no direct measurements of the offshore sediment plumes were made at the time of the cyclone, anecdotal evidence from fishers and recreational divers suggests that, in the months after Bola, much of the seabed was mantled by a layer of fluid mud (Bennett 1989). Ecological surveys conducted by scuba divers in 1992 revealed that most of the Bola mud layer had disappeared from reef areas on the inner shelf, leaving benthic communities that were depleted in species numbers and diversity (Battershill 1993).

In this paper we seek to reconstruct the offshore dispersal and deposition of mud sediment under normal and Cyclone Bola-type events using a mixture of historical and observational data. The latter includes a series of high-resolution, 3.5 kHz seismic reflection profiles, the uppermost layers of which are resolved by gravity cores (Fig. 2). The cores were analysed routinely for sedimentary structures and grain size. Together with the seismic reflection data, they provide a geographic and historic perspective of mud deposition as well as an insight into the processes affecting the mud depocentres. Mud dispersal was evaluated further from available satellite imagery, combined with physical oceanographic data.

ENVIRONMENTAL FRAMEWORK

General setting

The area of study comprises Poverty Bay and the adjacent continental shelf off eastern North Island (Fig. 1). There, the shelf extends 65 km between Mahia Peninsula and Monowai Rocks, extending 22–26 km seaward to the shelf break at 140–170 m depth. The normally gently inclined seabed is disrupted by several reefs that are aligned roughly parallel with the shelf edge. These features rise from depths of 70 m to break the surface at Ariel and Penguin Rocks, which collectively comprise Ariel Bank (Fig. 1).

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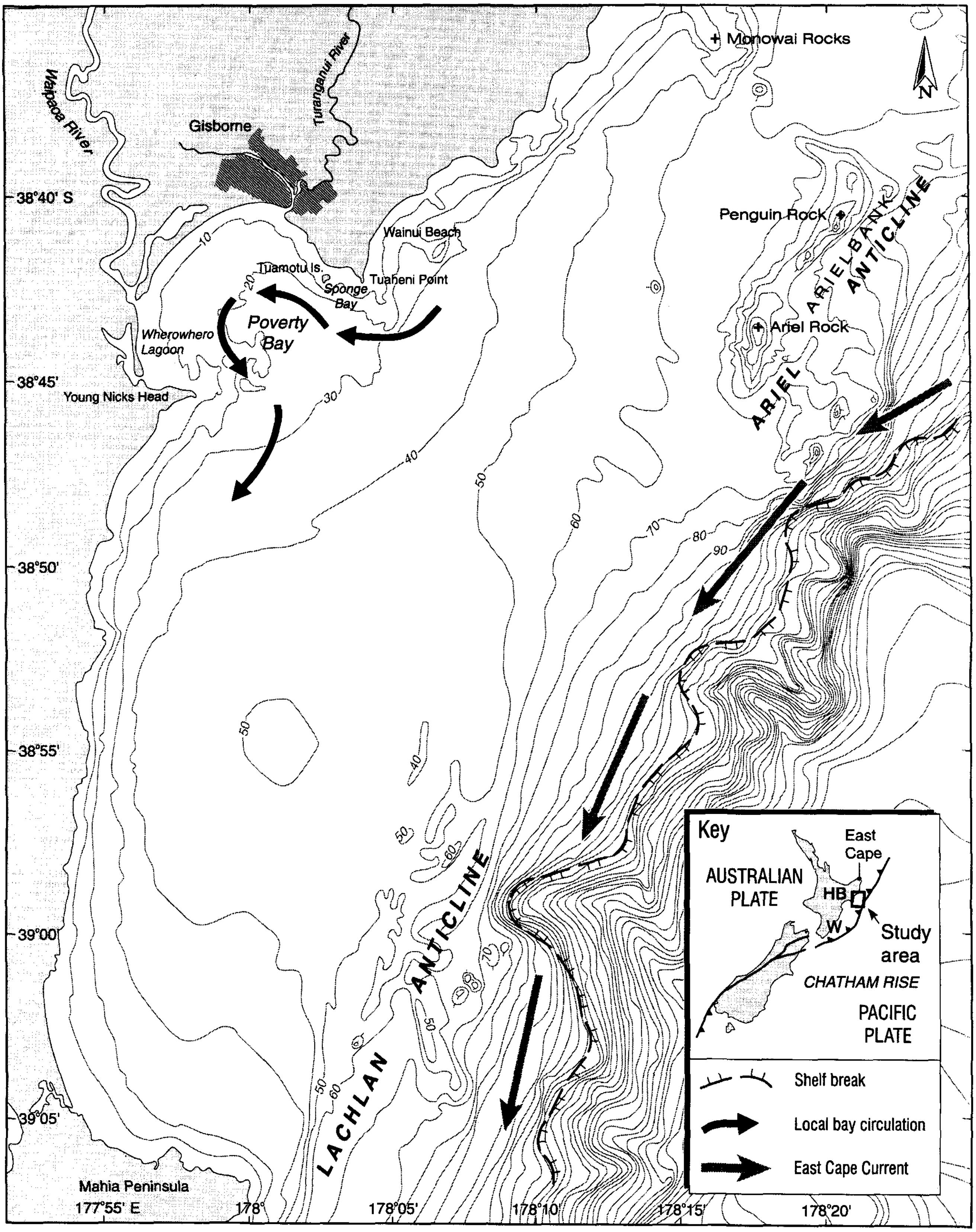


Fig. 1 Poverty Bay and the eastern continental margin showing metric bathymetry derived from Arron & Lewis (1992) and the main elements of the shelf circulation including the East Cape Current (Heath 1985) and Poverty Bay currents (Miller 1981). *Inset*: The New Zealand plate boundary.

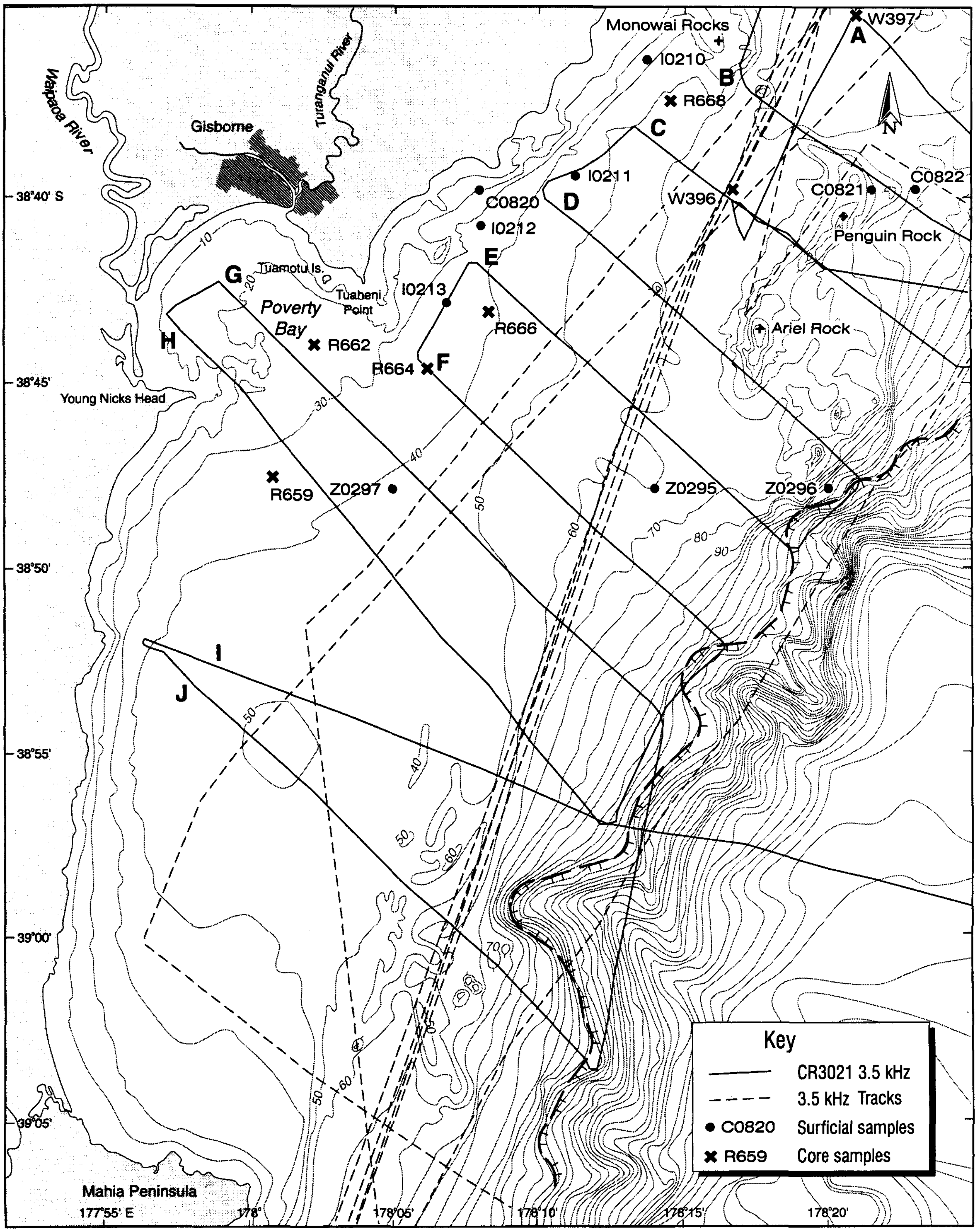


Fig. 2 Core and surface sample locations together with 3.5 kHz seismic profile tracks.

Poverty Bay is exposed to the southeast whilst being protected by Young Nicks Head to the south and Tuaheni Point to the northeast. The bay is 8.5 km wide at its entrance and is bordered by 13.4 km of mainly sandy beaches interrupted by the mouths of Wherowhero Lagoon and Waipaoa River (Smith 1988). The northeastern end of the sandy beach system ends near the Turanganui River mouth adjacent to Gisborne Harbour.

Tectonic framework

The Poverty Bay continental shelf lies within a zone of active deformation associated with the Hikurangi subduction margin (Fig. 1, inset). More precisely, the shelf straddles the top of the Neogene accretionary prism and the forearc basin, as outlined by Lewis (1980) and Lewis & Pettinga (1993). The boundary between prism and basin extends along the outer continental shelf and is defined by the landward limit of imbricate thrust ridges (Lewis & Pettinga 1993). Off Poverty Bay, two deep seismic reflection profiles outline a thrust ridge which is expressed as Ariel Bank—a tight anticline with a core of Miocene mudstone (Katz 1975). Landward of Ariel Bank, a synclinal basin filled with Pliocene–Pleistocene sediments has developed, but its full extent is not apparent because of the sparse survey coverage (Katz 1975; Lewis 1980). To the south of Poverty Bay, high-resolution 3.5 kHz profiles collected during the course of the present study identify another emergent fold that may be a continuation of the Lachlan Anticline (Lewis 1971, 1973).

Climate

Climate is a key factor controlling the input of sediment to the Poverty Bay shelf. The average annual rainfall for the 1937–80 period at Gisborne Airport was 1058 mm/yr (Hessell 1980; New Zealand Meteorological Service 1980). This value, however, hides the marked variability that occurs on a monthly and annual basis. Months of drought are punctuated by periods of intense rainfall. Deluges usually coincide with the passage of tropical cyclones which pass within 100 km of New Zealand once or twice a year (Sinclair 1993). There is also considerable variability in cyclonic intensity.

Cyclones aside, the region is not windy. The mean wind speed for Gisborne is <11 km/h, and gale-force gusts (>63 km/h) are recorded on only 48 days per year on average (Hessell 1980). Gusts >96 km/h blow only once or twice annually and may also be associated with tropical cyclones. The predominant wind is from the northwest, but it often weakens in the afternoon and is replaced by sea breezes generated from the southeast. In addition, the shelf is subject periodically to gales and storms from the south.

Hydraulic regime

Waves

The prevailing wave climate along much of the New Zealand east coast is dominated by easterly to southerly waves, with the major long-period swell component being from the south (Pickrill & Mitchell 1979). Southerly swell along this coast generally has height, H, of 0.5–2.0 m and period, T, of 6–9 s. Storm waves of up to 6 m have been recorded at East Cape, but rarely exceed 3 m in deep water (Pickrill & Mitchell 1979; Harris et al. 1983).

Coastal observations from within Poverty Bay (Smith 1988) confirm the prevalence of waves from the south and southeast, and identify average wave heights of 0.8–0.9 m and periods of 9–10 s. The wave climate, however, exhibits some variability but with no obvious seasonality. During local storms, for example, wave heights may exceed 4 m, whereas swell, generated at distant southerly storm centres, may reach 6 m (Smith 1968). Miller (1981) reported 18 months of data from a wave-rider buoy in Poverty Bay which, when combined with observations made by the Gisborne Harbour Board for the same interval, yielded a mean significant wave height, H_s, of 1.04 m and a mean zero up-crossing period, T_z, of 8.7 s. Seasonal variability was also noted within the dataset. Winter months had higher H_s and T_z values of 1.12 m and 8.96 s, respectively, whereas summer H_s and T_z values were lower at 0.88 m and 8.33 s, respectively.

Currents

Within Poverty Bay the surface circulation comprises an anticlockwise gyre that may reverse periodically under southeasterly winds at least near Gisborne Harbour (Williams 1966). The anticlockwise circulation is supported by satellite imagery of sediment plumes (Miller 1981; this study). The few short-term measurements of the bay currents suggest that they are slow (Miller 1981; Kensington 1990). The tidal component, for example, reaches only c. 10 cm/s.

The largest feature of the shelf circulation is the semi-permanent East Cape Current (Heath 1985; Chiswell 1994). This warm, saline flow passes south over the outer shelf and upper continental slope between East Cape (37°S) and Wairarapa (41°S) before turning eastward along Chatham Rise. At times the current reverses locally—a change that is caused by meanders in the flow (Denham et al. 1984) and/or southerly winds. The only direct measurements of current speed were made in near-surface waters off East Cape (Heath 1980). There, a velocity of 18 cm/s clockwise around the Cape was measured for a 10 day period.

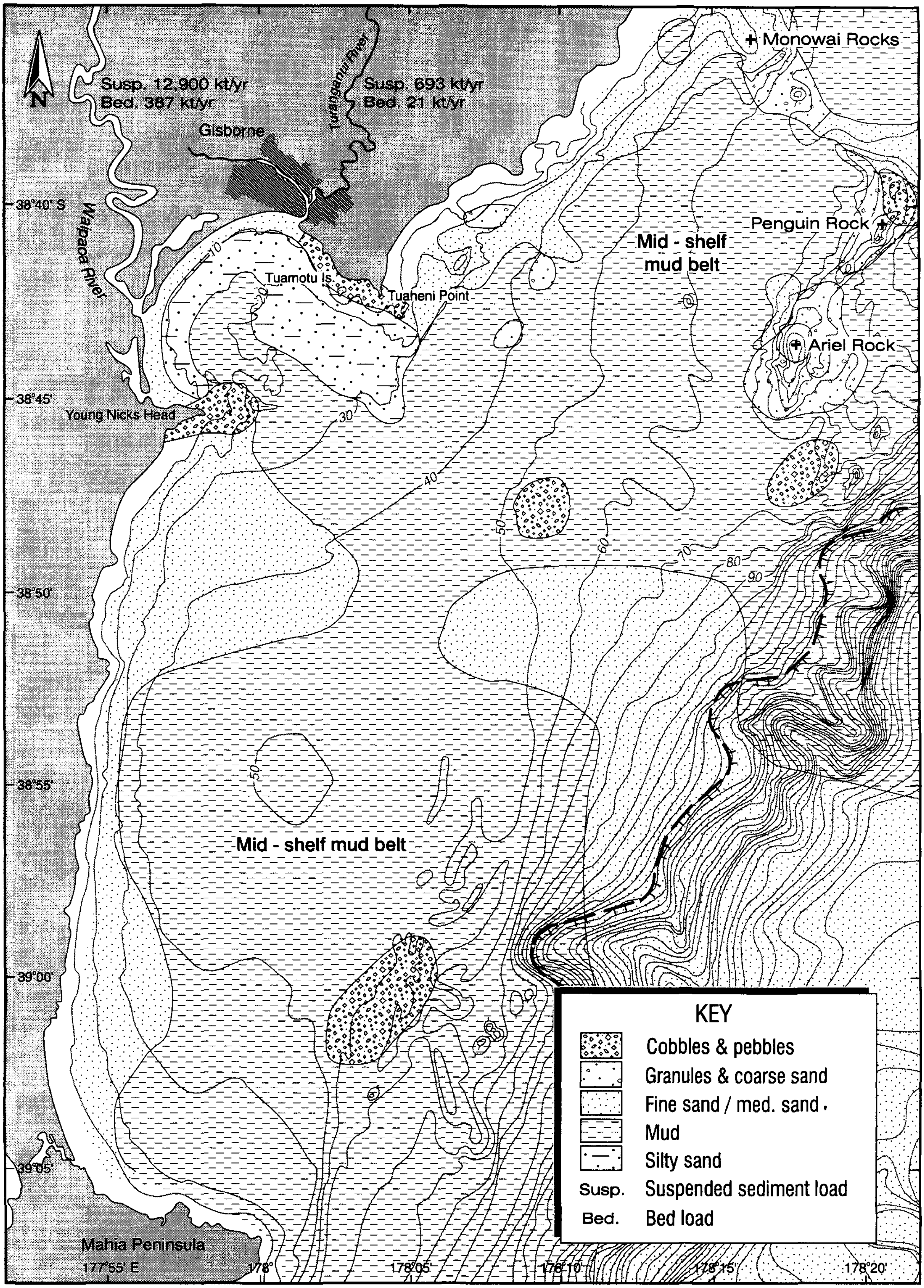
Inshore of the East Cape Current, there may be periodic flows to the north, with the Southland Current extending to southern Hawke Bay (Heath 1975) and possibly farther north under forcing conditions. Southerly winds set up a northerly flow, as attested by the drift-card data of Brodie (1960) and dispersal patterns of sediment plumes (see section on Mud Dispersal).

The tidal component of the open shelf circulation is usually weak (Carter & Heath 1975). A few hydrographic observations note that the tides are "not felt" >9 km offshore and that their direction is readily influenced by the wind (British Admiralty 1987).

Sediments

The floor of Poverty Bay is mantled by sand and mud, the mean grain size becoming progressively finer offshore (Fig. 3). Sediments range from coarse sand near river mouths and beaches, to silty fine sand in the central bay, and mud (Pantin & Gibb 1968; Miller 1981). The bulk of these

Fig. 3 Distribution of surficial sediments within Poverty Bay (modified from Miller 1981) and on the adjacent continental shelf (modified from Pantin & Gibb 1968). Annual suspended and bedload contributions for the Waipaoa and Turanganui Rivers are presented. ➤


Susp. 12,900 kt/yr
Bed. 387 kt/yr
Susp. 693 kt/yr
Bed. 21 kt/yr
Turanganui River
Gisborne
Waipaoa River
Monowai Rocks
Penguin Rock
Ariel Rock
Mid - shelf mud belt
Tuamotu Is.
Tuaheni Point
Young Nicks Head
Mid - shelf mud belt
Mahia Peninsula
38°40' S
38°45'
38°50'
38°55'
39°00'
39°05'
177°55' E
178°
178°05'
178°10'
178°15'
178°20'
KEY
Cobbles & pebbles
Granules & coarse sand
Fine sand / med. sand
Mud
Silty sand
Susp. Suspended sediment load
Bed. Bed load

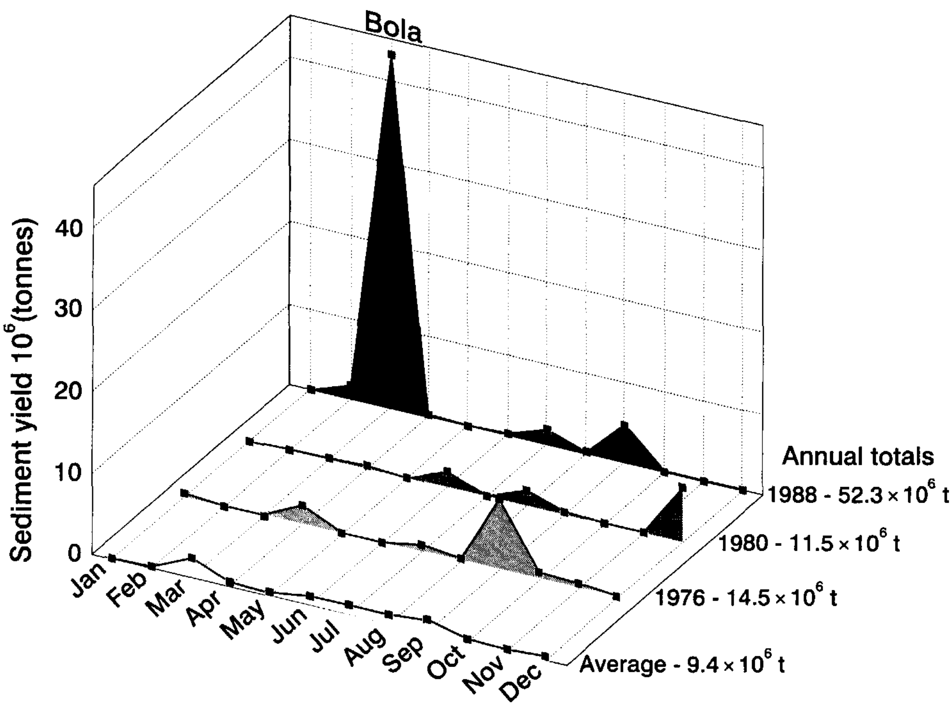


Fig. 4 Monthly suspended sediment yields (load) for the Waipaoa River for an average year and years of major cyclones, including 1988 when the 100 year storm of Cyclone Bola struck the North Island.

deposits is derived from the Waipaoa River (Miller 1981; Kensington 1990; Sanders 1993; Wood 1993). This river is the main point-source of sediment for the region, and its impact is manifested by a cross-shelf lobe of mud extending seawards from the Waipaoa River mouth (Fig. 3).

On the open shelf, outside Poverty Bay, the distribution of surficial sediments follows the classic pattern for a modern depositional setting (e.g., Carter 1975). Sands occupy the inner shelf and are gradually replaced by mud at 30–40 m depth. The modern mud blanket extends to the shelf edge except in the vicinity of the anticlines on the outer shelf. Here, exposures of Neogene sedimentary rocks (Lewis 1973; Katz 1975) are surrounded by aprons of gravelly (cobbles, pebbles) and sandy sediments (Fig. 3). Another exception is the gap between the Lachlan and Ariel Anticlines, which is floored with fine sand.

SEDIMENT INPUT TO POVERTY BAY

Coastal erosion

Sediment supplied by coastal erosion to Poverty Bay is small, since much of the coastline has prograded at an average rate of 0.2–0.8 m/yr for the last 2000 years (Pullar & Penhale 1970; Smith 1988). Progradation accounts for deposition of c. 3.97 × 10^6 m^3 of sediment. However, localised erosion has been noted, mainly at Tuaheni Point and Sponge Bay (Gibb 1978).

Outside of Poverty Bay, coastal growth is variable (Gibb 1978). The northern reach of Wainui Beach, for example, has been subject to both erosion (–0.2 m/yr) and progradation (+0.13 m/yr) between 1887 and 1976.

Fluvial input

The dominant sediment source is the Waipaoa River, which delivers an estimated annual total load of 12.9 × 10^6 t/yr to the ocean (Griffiths & Glasby 1985). This figure is corroborated by measurements made by the Gisborne District Council (unpubl. rep.), which recorded 9.4 × 10^6 t/yr. Such a supply is the fourth largest for any river in New Zealand and is generated from a catchment area with the fifth highest specific sediment yield at 5836 t/km^2 per year (Griffiths & Glasby 1985). By comparison, the annual discharge of the nearby Turanganui River is only c. 0.69 × 10^6 t/yr (Gisborne District Council unpubl. report).

Because of the prevalence of soft, fine-grained rocks in the river catchments around East Cape, c. 97% of the total sediment load is suspended material, composed of mud and fine sand (Adams 1980; Miller 1981; Griffiths & Glasby 1985).

The amount of suspended load fluctuates markedly depending upon the amount and intensity of rainfall in the catchment area. From a seasonal perspective, loads are smallest in the summer (November–February) when the monthly average is >0.2 × 10^6 t. By comparison, winter (June–September) discharges are >1.25 × 10^6 t. Superimposed on this seasonality are periodic floods of sediment associated with tropical cyclones and storms which usually arise in the Southwest Pacific between December and April (Fig. 4).

Sediment input from Cyclone Bola, March 1988

The most recent and severe tropical cyclone to reach the Poverty Bay region was Cyclone Bola on 5–10 March 1988. This disturbance moved across northern New Zealand and then southwards, causing strong easterly winds up to severe gale force (c. 100 km/h). The cyclone incorporated moist tropical air which condensed and cooled to form a thick cloud sheet over the East Cape/Gisborne region (Prasad 1988; Service de la Meteorologie 1988; Singleton et al. 1989a, b; Sinclair 1993). The resultant heavy precipitation was the highest ever recorded for the North Island. Up to 900 mm

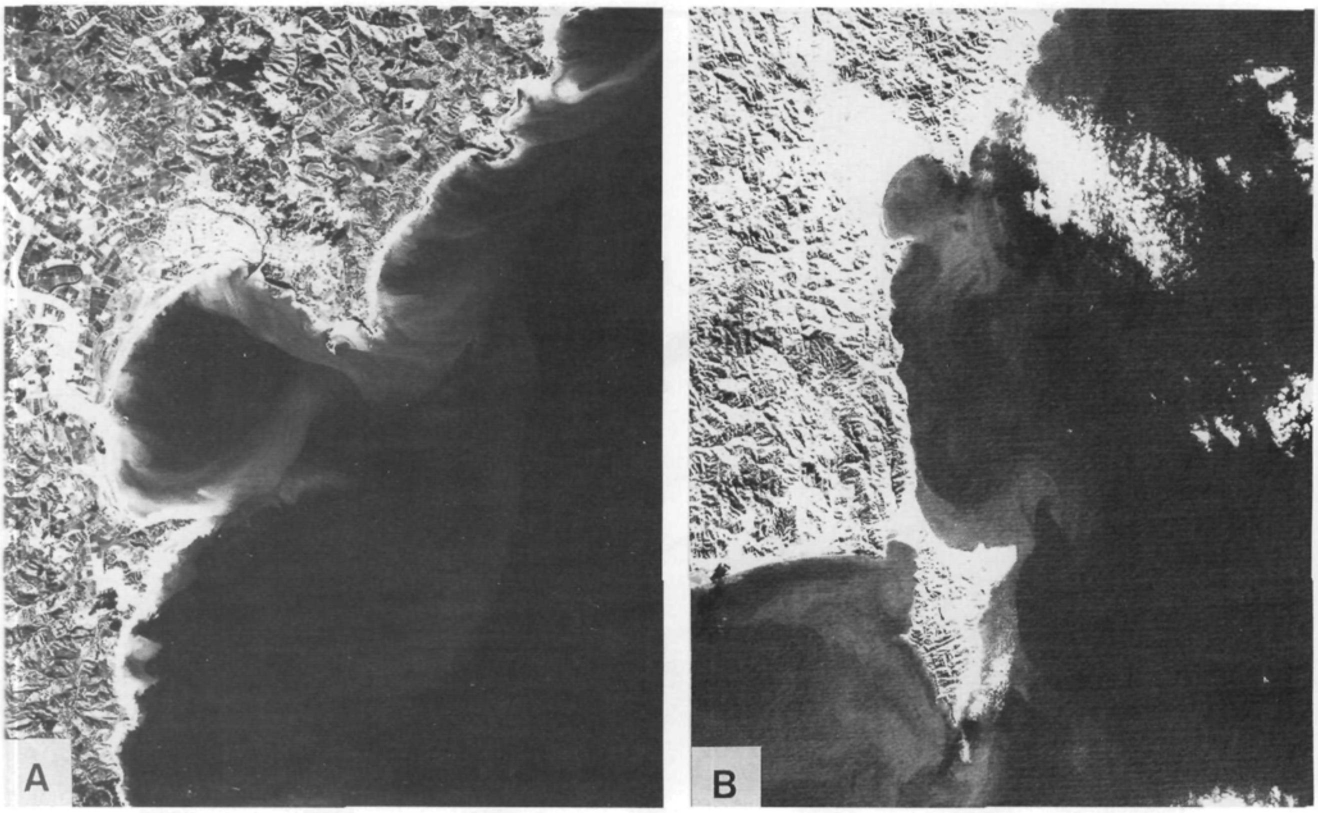


Fig. 5 SPOT (A) and Landsat (B) satellite images highlighting the northward and southward transport of sediment plumes, respectively. (SPOT Image (CNES) supplied by Landcare Research.)

of rain fell on the East Cape and Hawke's Bay regions in 72 h (Singleton et al. 1989a, b; Sinclair 1993). Around East Cape the heaviest rainfall was over steep hill country composed of highly erodible, soft Tertiary siltstones and mudstones (Singleton et al. 1989a). Severe erosion caused river systems to aggrade rapidly, resulting in flooding of surrounding areas (Singleton et al. 1989a). Trotter (1988) estimated, from analysis of satellite imagery, that 10–20% of the hill country in the East Coast–Gisborne region had undergone severe landsliding in response to Cyclone Bola. As a result, Waipaoa River registered its highest flow of 5300 m^3/s, the highest on record, compared to the average flow of 38 m^3/s (Gisborne District Council 1994).

The suspended sediment load for the 6 day Bola event exceeded 40×10^6 t or more than five times the mean annual suspended load (Fig. 4). Kensington (1990) reported infilling of the Gisborne Harbour entrance off northern Poverty Bay, immediately after Cyclone Bola, at a rate of 2426 m^3/day as compared to the average rate of only 107 m^3/day. Sediment accretion was also noted by Brown (1995) to occur along Wainui Beach, but by 1990 this had reverted back to a delicate balance between progradation and erosion.

MUD DISPERSAL

When the Waipaoa River discharges into Poverty Bay, the mud component of the suspended load is contained in a buoyant or hypopycnal plume, the dispersal of which can be observed on satellite images. Landsat, Coastal Zone Colour Scanner, and SPOT images (Fig. 5) show that the main part of the Waipaoa plume moves anticlockwise around the southern coast of Poverty Bay and out onto the open shelf. A similar, topographically influenced dispersal occurs about the northern shoreline of Mahia Peninsula (Miller 1981). Linear transport of the plume, however, is complicated by the occurrence of transient eddies from promontories such as Young Nicks Head and Mahia Peninsula. Once on the middle shelf, beyond the confines of the embayment, surface plumes travel mainly along the shelf to either the northeast or south, depending upon the wind direction. As a plume spreads onto the outer shelf it comes under the influence of the East Cape Current, and therefore in this zone a net southward transport is expected (e.g., Stevenson et al. 1977). Occasional northward incursions may occur, however, as the East Cape Current reverses, presumably in response to meanders and/or wind forcing (British Admiralty 1987; Chiswell 1994). Given the net flow southwards, the local mud budget on the outer shelf may be supplemented by suspended load from rivers discharging north of the study area.

At times, concentrations of fluvial sediment may be sufficiently high to raise the density of the river discharge above that of coastal waters. Under these conditions, a negatively buoyant or hyperpycnal plume may form and extend over the shelf floor. The classic example of this phenomenon is Huanghe River, in the People's Republic of China, where sediment concentrations of 25–220 kg/m^3 are sufficient for the sustained generation of hyperpycnal plumes (Wright et al. 1986; Wright 1989). In Waipaoa River, sediment concentrations reach c. 26 kg/m^3 during annual floods, c. 40 kg/m^3 in the 10 year event, and a recorded maximum of 58 kg/m^3 during Cyclone Bola (unpubl. data

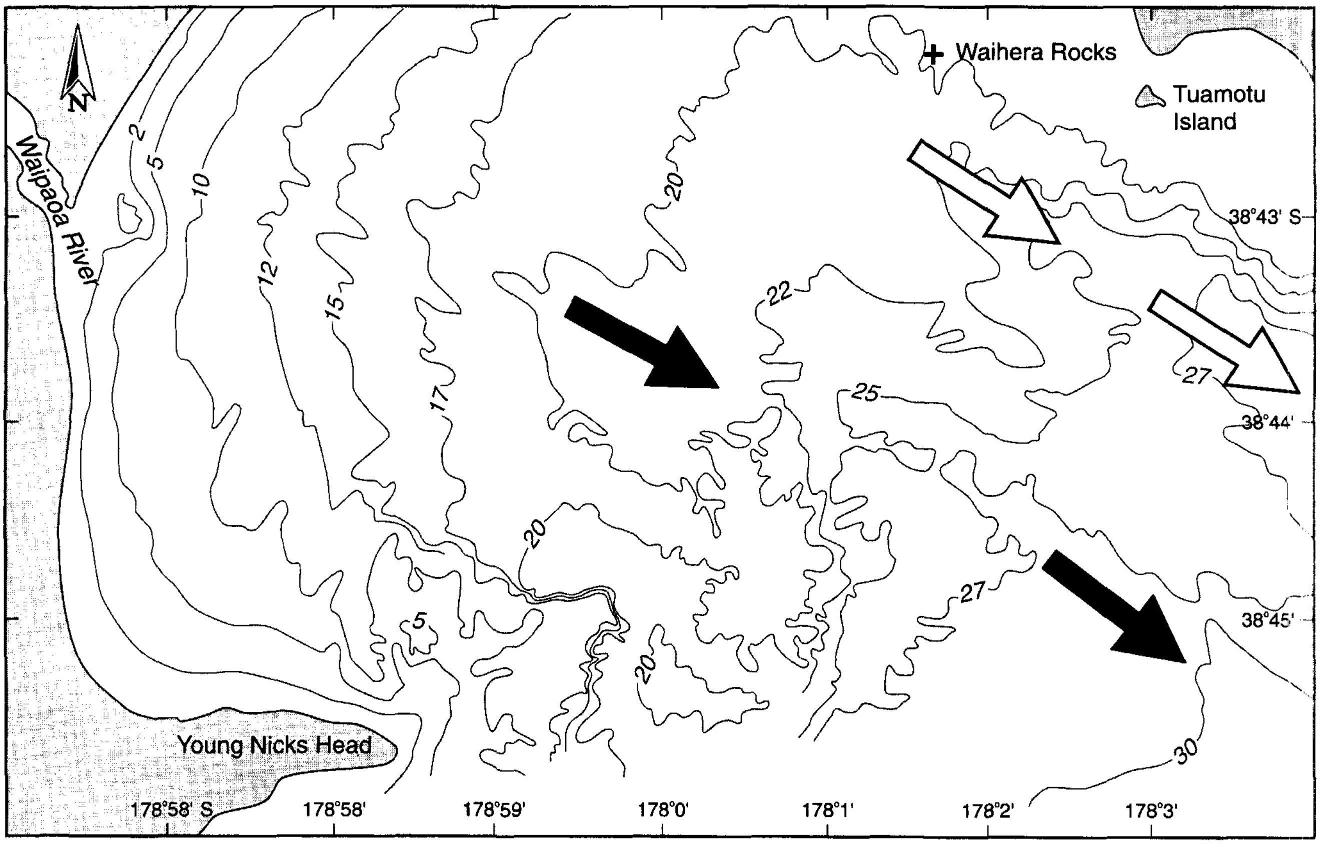


Fig. 6 Detail of metric bathymetry in southern Poverty Bay outlining the broad trough (solid arrows) leading from the Waipaoa River. The trough in the north of the bay (open arrows) is a relic from when the Waipaoa River discharged in that reach.

from NIWA Water Resources Databank and Gisborne District Council).

Assuming a density of local coastal waters (ρ_{cw}) of 1.0250–1.0260 g/cm^3 as derived from local temperature and salinity data (Garner 1961; Ridgway & Stanton 1969), then the amount of river sediment needed to raise the discharge density (ρ_{rw}) above that of the coastal waters is 42–43 kg/m^3 (Gilbert 1983). The sediment rating curve for Waipaoa River (unpubl. data from NIWA Water Resources Databank) shows this threshold concentration is reached when the flow exceeds 2000 m^3/s. According to Singleton et al. (1989b), flows of that magnitude have a return frequency of 10 years. Thus, assuming that mixing will not alter the relationship whereby $\rho_{rw} > \rho_{cw}$, then conditions will favour hyperpycnal flows on a decadal frequency.

The movement of subsurface turbid water, whether it is contained within a hyperpycnal plume or within a simple bottom nepheloid layer, is at the mercy of the shelf hydraulic regime and of any momentum of the plume itself. With regard to the latter, dense hyperpycnal flows may act as gravity flows, moving downslope along self-eroded or pre-existing channels (e.g., Wright et al. 1986). The existence of a shallow trough extending southeast from off the Waipaoa River mouth to at least 35 m depth, may be one such plume pathway (Fig. 6). Although the Holocene sediments are thinner in the trough compared to the surrounding seabed, seismic data are inconclusive as to whether the trough is being eroded actively or is simply an abandoned river channel guiding plumes seaward. Outside the trough, basal plumes probably move in accord with the prevailing circulation. Under northwest to north winds, Ekman forcing of the wind-drift current (to the left of the wind direction in the Southern Hemisphere) will cause the offshore transport of surface waters, which is compensated by shoreward movement of bottom waters—an upwelling circulation (Heath 1972). As a result, any existing turbid bottom waters on the middle and outer shelf may move shorewards. The reverse may occur under southerly conditions. Surface waters and the attendant sediment plumes may move shoreward; this motion is compensated by the offshore transport of the bottom waters (i.e. downwelling). Accompanying currents can reach 10 cm/s or more and are thus a significant component of the shelf circulation (e.g., Smith 1981; Nittrouer & Wright 1994). Near the outer shelf, hyperpycnal flows and benthic nepheloid layers are liable to be constrained by the rising anticlines, except in the vicinity of the gap off Poverty Bay.

In addition, sediment plumes form through wave-induced suspension of nearshore bottom sediments and through coastal erosion of soft, Tertiary sediment outcrops (Miller 1981). On satellite images (Fig. 5), coastal plumes usually extend alongshore but may locally project seaward in zones of probable rip currents.

MUD DEPOSITION

Fluid mud

When sediment concentrations in the benthic nepheloid layer reach 5–10 kg/m^3, as is probably the case during the annual

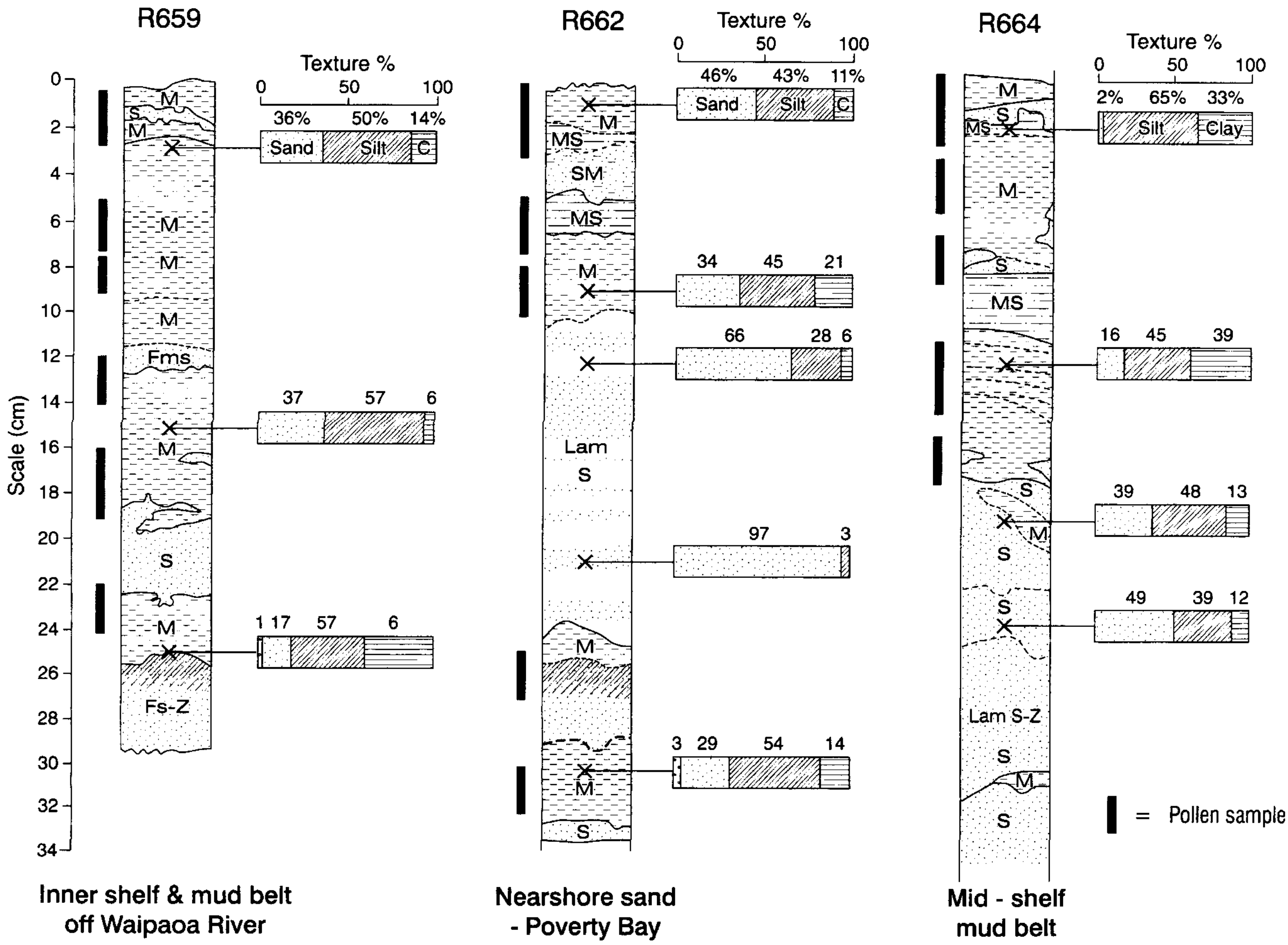


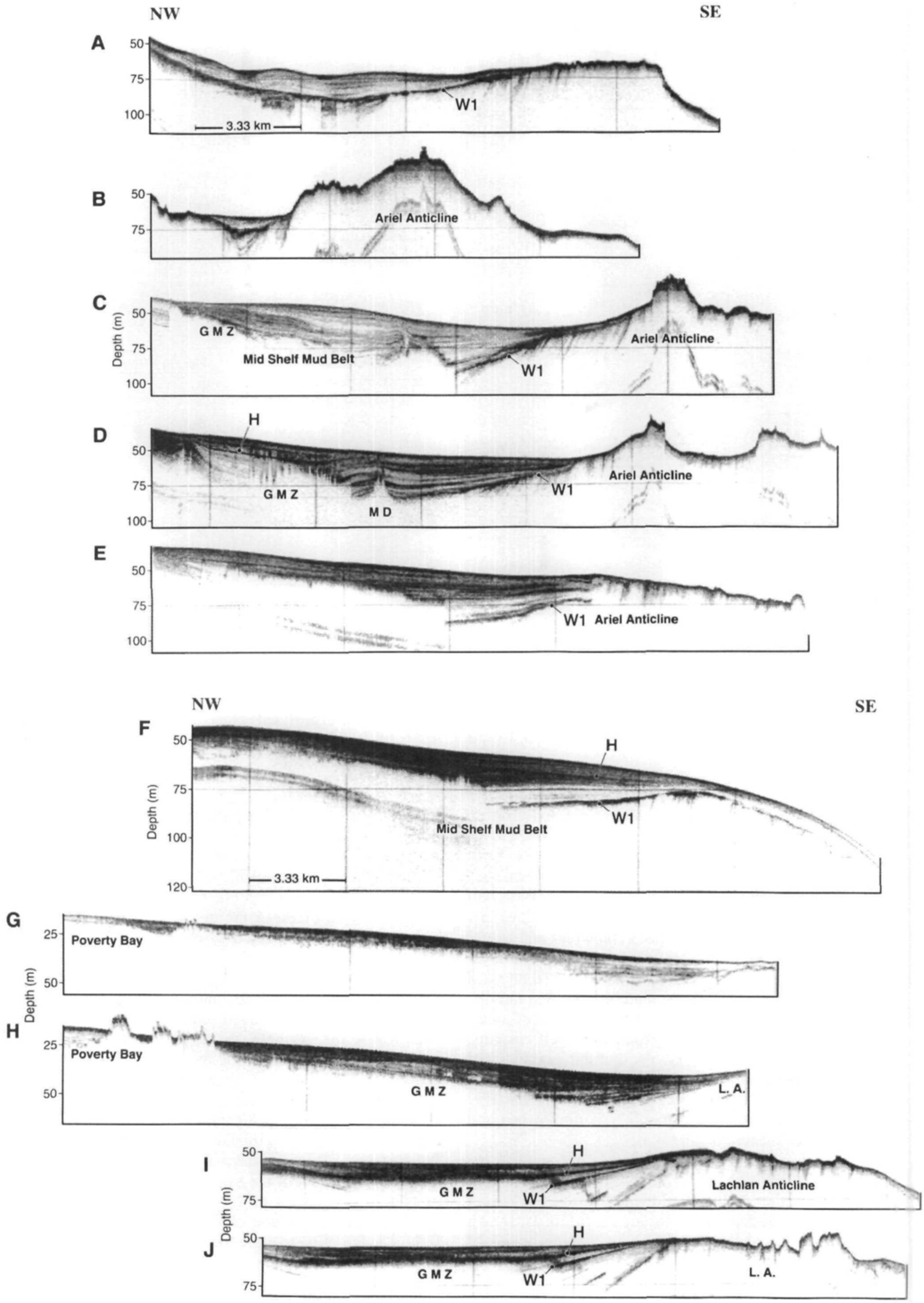
Fig. 7 Short cores from the inner to middle shelf near Poverty Bay displaying the textural variability associated with the interfingering of the nearshore sand prism and the middle shelf mud blanket—a consequence of high mud supply and deposition following river floods punctuated by periods dominated by shelf current/wave processes. Locations of analysed pollen samples are depicted, and core locations are presented on Fig. 2. S, sand; Fs, fine sand; Fms, fine muddy sand; SM, sandy mud; Z, silt; M, mud.

flood of Waipaoa River, settling mud particles may start to become cohesive and form a high-concentration suspension or "fluid mud" layer above the bed (McCave 1984; Dyer 1994). Observations by divers and fishers suggest that such a layer extended over the inner to middle Poverty Bay shelf after the 1988 Bola flood. Estimates of the layer thickness ranged up to 2 m as measured against a shipwreck and lobster pot lines located near reefs throughout the bay. The layer also smothered the resident benthic communities (Battershill 1993). Local reports also suggest that the layer was mobile, as attested by the covering and uncovering of reef areas. Such mobility was probably in response to shelf currents, although fluid mud layers may also exhibit gravitational flow. Once mobile, mud layers may be dispersed by currents (Dyer 1994). Alternatively, layer cohesiveness and viscosity can increase to a point where mobility is diminished and deposition takes place.

Modern mud blanket

As reported for other shelves worldwide (Nittrouer & Wright 1994), the principal depocentre for mud is the middle reach of the Poverty Bay shelf (Fig. 3). The nearshore boundary (>50% mud) of the Poverty Bay mud blanket is in depths of 30–45 m on the open shelf, but shallows to 15 m depth within the bay itself, off the Waipaoa River mouth (Fig. 3). The position of this mud boundary is mainly a function of the interaction between mud supply and the erosional capacity of the prevailing flow which, on the inner shelf, is dominated by swell and wind-driven currents (McCave 1972; Drake 1976). As one of these factors prevails, the inner mud boundary can be expected to change. Analysis of short (<1 m) subsurface cores from near the boundary supports such variability (Fig. 7). Layers of mud, presumably deposited at times of high sediment supply relative to current competency, are interbedded with sand and silty sand. In core R664, for example, the sand layers may represent periodic seaward shifts of the mud boundary, as originally charted by Miller (1981), although the general increase in mud upcore suggests the overall displacement of the boundary is landward. This possible change occurred over a near-flat seafloor proximal to the Waipaoa River. North of Poverty Bay, away from the main sediment supply and over a more inclined shelf, the sand/mud boundary may be less mobile.

Given the magnitude of fluvial input during Cyclone Bola, it is tempting to correlate the uppermost 10–18 cm of mud-dominant core sediment with that 1988 event (Fig. 7). We have been unable to substantiate the precise age of this


NW
SE
A
50
75
100
W1
3.33 km
B
50
75
Ariel Anticline
C
Depth (m)
50
75
100
G M Z
Mid Shelf Mud Belt
W1
Ariel Anticline
D
H
50
75
100
G M Z
M D
W1
Ariel Anticline
E
50
75
100
W1
Ariel Anticline
NW
SE
F
Depth (m)
50
75
100
120
H
W1
Mid Shelf Mud Belt
3.33 km
G
Depth (m)
25
50
Poverty Bay
H
25
50
Poverty Bay
G M Z
L. A.
I
Depth (m)
50
75
H
W1
G M Z
Lachlan Anticline
J
50
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muddy layer by radiometric dating owing to the geological "youth" of the deposit. Instead, palynology was used to find a modern time plane, c. A.D. 1950, when *Pinus radiata* pollen was introduced to the Gisborne region (M. McGlone pers. comm. 1994). The uppermost 2–3 cm of core contain pine pollen with bracken and grass as subsidiary types. This assemblage equates well with the present pollen rain. In contrast, muds lower in the cores have markedly reduced pollen numbers and include a significant amount of reworked Tertiary fossil types (e.g., *Nothofagus brassii*) in addition to the pollen noted previously. This reduction in numbers may simply reflect poor pollen preservation. Alternatively, it may represent a major influx of mud eroded from older, pollen-deficient sediments—a situation that is consistent with the extensive erosion of the Tertiary terrain that accompanied the Bola event (e.g., Trotter 1988; Phillips 1989). The seaward border of the mid-shelf mud belt approximates the 70 m isobath north of Poverty Bay and the 50 m isobath to the south. On the surficial sediment chart (Fig. 3), this boundary follows approximately the discontinuous line of reefs and their attendant aprons of sand and gravel. Mud has also been deposited on the outer shelf, seaward of the reefs, but seismic profiles show that the main local depocentre is the middle shelf (Fig. 8).

Seismic reflection definition of mud depocentres

The suite of 3.5 kHz seismic profiles (lines A–J on Fig. 2) not only identifies the morphology of the mud deposits and the principal depocentre, but highlights the influence of tectonism on local sedimentation.

In essence, mud is accumulating in a broad depositional syncline situated mainly on the middle shelf in water depths of 30–70 m (Fig. 8 A–J). The synclinal axis extends at least 50 km southwest along the shelf, from Monowai Rocks to near Mahia Peninsula, where the structure appears to extend towards the Mahia Syncline in Hawke Bay (Lewis 1971). The landward flank of the syncline is marked by the rising coast where Holocene uplift rates of up to 4 m per 1000 years have been recorded (Ota et al. 1987). On the outer shelf the syncline is bordered by growing anticlines that are manifested in the bathymetry by Ariel Bank and unnamed shoals just north of Mahia Peninsula (Fig. 8). The latter shoals appear to be an extension of the Lachlan Anticline and are provisionally correlated with that structure (Lewis 1971, 1973). The other structure to the north is here informally termed the Ariel Anticline. The crests of Ariel and Lachlan Anticlines are offset directly opposite Poverty Bay, resulting in a 13 km wide gap in the anticlinal "barrier". This gap appears to be the extension of a southeast-trending structural depression that encompasses the lower course of the Waipaoa River (Brown 1995). Similar offsets of anticlines have been noted elsewhere on the eastern North Island shelf (Lewis 1971; Ota et al. 1987). Some offsets may be fractures arising from the collision and subduction of seamounts with the accretionary prism, such as is proposed to occur off Mahia Peninsula by Collot et al. (1995).

Fig. 8 3.5 kHz seismic profiles outlining the middle shelf mud belt within a growing syncline that is flanked in the southeast by equally active Ariel and Lachlan (LA) Anticlines. W1 is the unconformity at the base of the postglacial transgressive sequence; H is a Holocene reflector of c. 8000 yr B.P. GMZ = gaseous mud zone; MD = mud diapir. Profiles are located on Fig. 2.

Within the shelf syncline, prisms of sediment exhibit moderate acoustical transparency with 3.5 kHz seismic penetration down to 37 m. Thin, continuous reflectors dip at low angle to the east but are upturned against the flanks of the growing anticlines on the outer shelf (Fig. 8A–C, H–J). Thus, the anticlines act as barriers to the seaward progradation of the mid-shelf mud prism, except in the gap opposite Poverty Bay where the prism has prograded to the shelf edge (Fig. 8F).

The full internal structure of the basin fill is masked by a zone of presumed gaseous mud occupying the central reach of the basin at sub-seabed depths of 8–15 m (Fig. 8C–E). Elsewhere, reflectors are pierced and disrupted by acoustically opaque bodies which may be gas-charged mud diapirs (Fig. 8D). Such bodies are likely to form where water-saturated sediments are incorporated into accretionary prisms, and they have been identified in local terrestrial and marine settings of the Poverty Bay region (Nelson & Healy 1984; Barber et al. 1986).

The synclinal fill rests on a prominent unconformity, termed W1 (Lewis 1973), which is the uppermost erosional surface in the seismic section and therefore is regarded as having formed during the last glacial lowering of sea level, 18 000 years ago (Carter et al. 1986). W1 reaches a maximum subsurface depth of 37 m before it is masked by gaseous sediments. A second stratigraphic marker is present in most profiles as a prominent, conformable reflector (H) in the upper part of the basinal fill (Fig. 8). Reflector H typically lies 8–15 m below the seabed (Fig. 9A), where it merges with the gaseous zone in the synclinal axis. Lewis (1973) tentatively ascribed an 8000 year age to reflector H on the basis of radiocarbon dating from just below a presumed exposure of the reflector in Hawke Bay (Pantin 1966).

Using the W1 reference level, the thickness of the postglacial basin fill can be estimated, given the limitations imposed by acoustic masking by gaseous sediments. The thickest succession of postglacial sediment (37 m recorded max.) is found below the deep reach of the middle shelf, east and northeast of Poverty Bay (Fig. 9B). A simple interpolation below the gaseous zone suggests W1 may have a thickness of 55 m or more. By comparison, the bathymetrically shallower part of the syncline, south of Poverty Bay, has a maximum recorded thickness of 25 m for the post-W1 fill (Fig. 9B).

Inside Poverty Bay, a 12 m thick sediment prism has formed on the inner shelf (Fig. 8H). In the northern half of the bay this sediment body is continuous with the basin fill on the middle shelf. In the southern bay, however, the prism is partly ponded by reefs extending from Young Nicks Head (Fig. 8H), and is thus partly separated from the middle shelf. The Poverty Bay prism rests on a dissected W1 surface that was presumably eroded by the Waipaoa and Turanganui Rivers during the last lowering of sea level. Paleochannels can be traced to the mouth of the bay (Fig. 6), but any seaward continuations are masked by gaseous sediments.

Rates of accumulation

Sediment volumes were calculated from isopachs (Fig. 9) using the "trapezoidal" method outlined in Bannister & Raymond (1972). The results were checked against those obtained from an ARC-INFO geographical information system and were found to agree. For the calculation it was

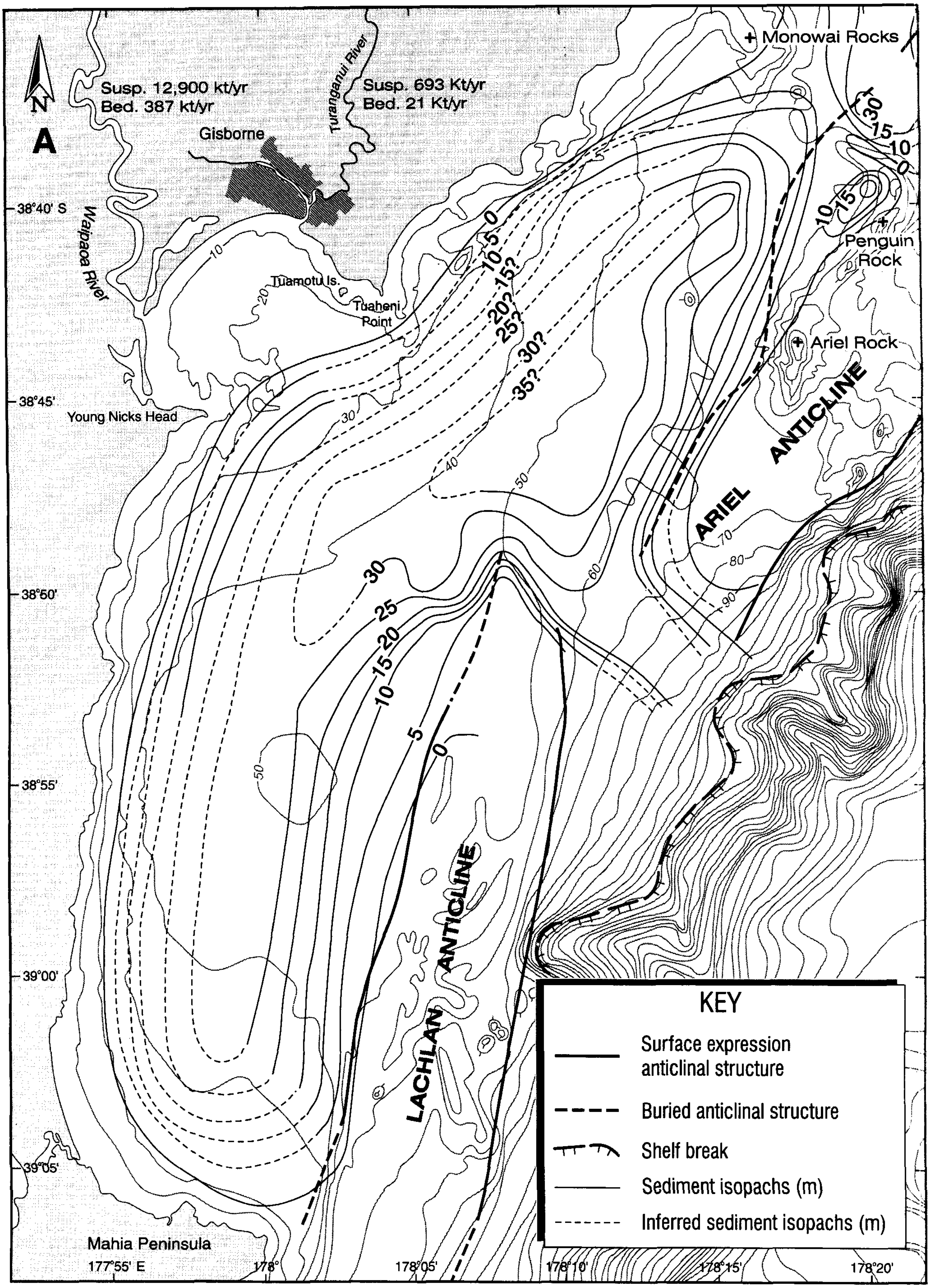
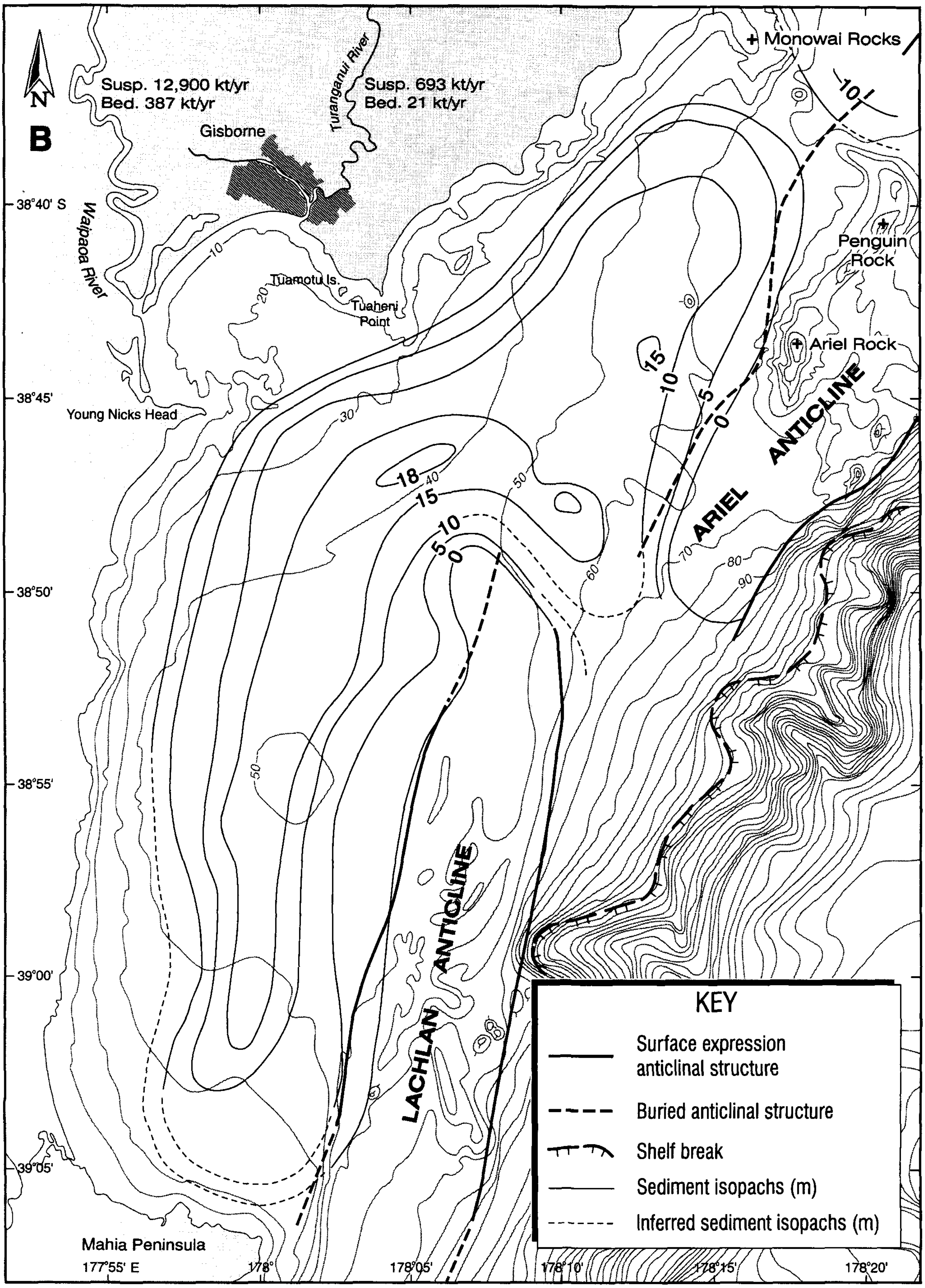


Fig. 9 (*above and opposite*). Isopachs to reflector W1 (**A**), and reflector H (**B**), outlining the main depocentre for postglacial mud



N
B
Susp. 12,900 kt/yr
Bed. 387 kt/yr
Turanganui River
Susp. 693 kt/yr
Bed. 21 kt/yr
Gisborne
Monowai Rocks
Penguin Rock
Ariel Rock
Waipaoa River
Tuamotu Is.
Tuaheni Point
Young Nicks Head
Mahia Peninsula
ARIEL ANTICLINE
LACHLAN ANTICLINE
38°40' S
38°45'
38°50'
38°55'
39°00'
39°05'
177°55' E
178°
178°05'
178°10'
178°15'
178°20'
KEY
Surface expression anticlinal structure
Buried anticlinal structure
Shelf break
Sediment isopachs (m)
Inferred sediment isopachs (m)

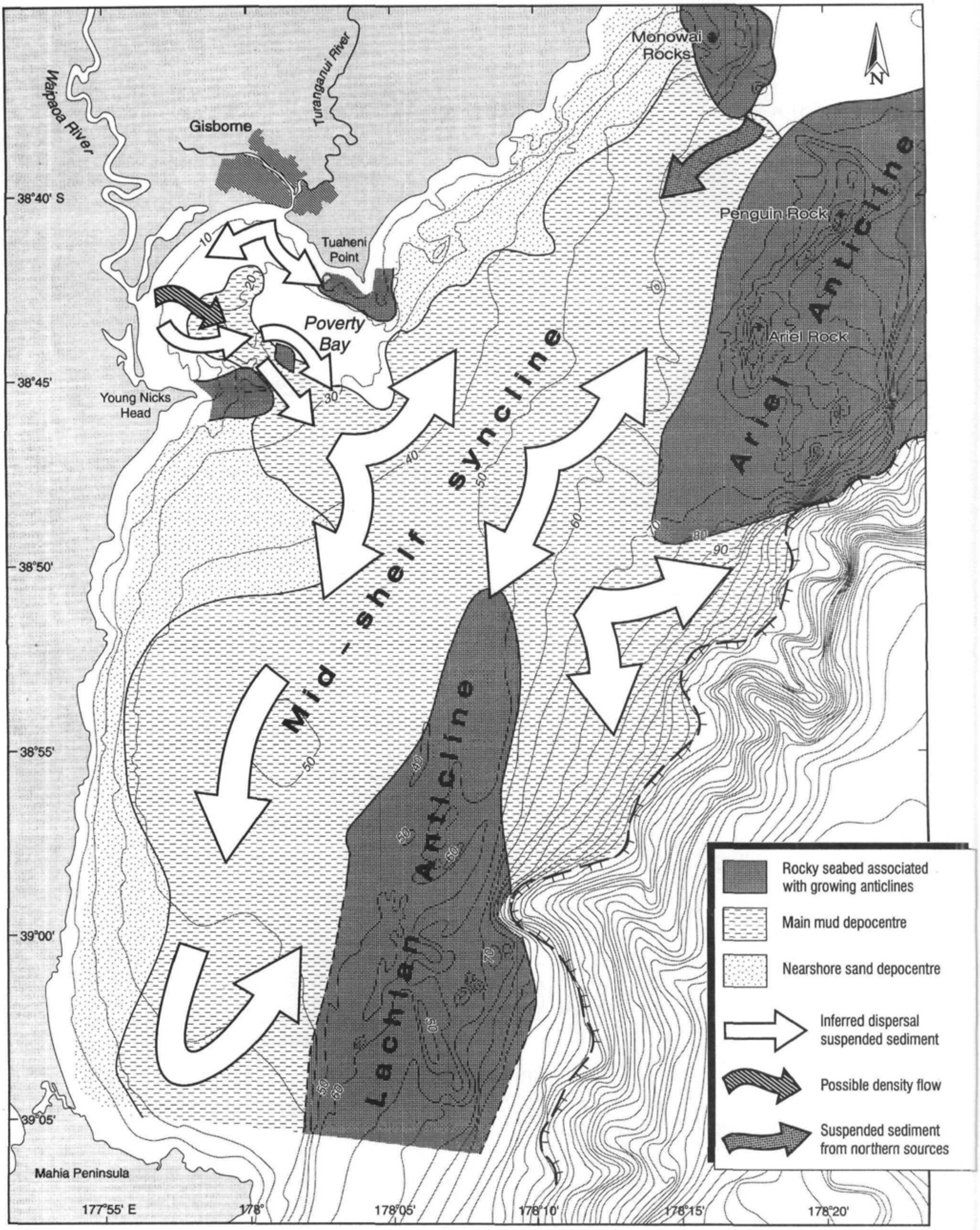


Fig. 10 A model for Poverty Bay mud sedimentation incorporating the structural elements that control the mud depocentre, the main dispersal paths, and the middle shelf mud depocentre.

assumed that the mid-shelf mud deposit had compacted by c. 15% of volume, as inferred from burial curves in Reike & Chilingarian (1974). To obtain sediment mass, a density of 1.45 t/m^3 was assumed on the strength of measurements made on similar lithologies on the nearby Hawke Bay shelf (Barnes 1991).

Between 18 000 yr B.P. and 8000 yr B.P., 12 km^3 of sediment accumulated in the mid-shelf mud belt at a rate of 1.2×10^6 m^3/yr or 1.7×10^6 t/yr. After 8000 yr B.P., the volume deposited decreased to 8 km^3, the accumulation rate being 1×10^6 m^3/yr or 1.45×10^6 t/yr. With respect to the modern input of mud to the Poverty Bay shelf, the Waipaoa and Turanganui Rivers deliver a total of 13.6×10^6 t/yr suspended load and 0.5×10^6 t/yr bedload (Griffiths & Glasby 1985). If all the bedload and the sand component of the suspended load are confined to the nearshore sand prism, which Smith (1988) noted had an accumulation rate of 5×10^6 t/yr in Poverty Bay, then c. 9×10^6 t/yr of mud reaches the middle shelf and beyond. The modern mud supply is, therefore, nearly five times more than the post-8000 yr B.P. accumulation rate of 1.45×10^6 t/yr.

This difference may reflect escape of mud from the shelf and/or a change in the rate of mud supply. With respect to the latter, terrestrial erosion in eastern North Island increased markedly in the late nineteenth century with the deforestation that accompanied European settlement (e.g., De Rose et al. 1993). Although the impact of this event on the nearby continental shelf has yet to be quantified, an idea of the magnitude of the change may be gained from cores collected in Lake Tutira in central Hawke's Bay. The data of Page et al. (1994) indicate post-European sedimentation rates of at least 14 mm/yr. In contrast, a 16 m long core described by Eden et al. (1993) from the same lake has an average pre-European rate of c. 2 mm/yr going back to c. 6500 yr B.P. Although generalisations based on one small lake catchment should be made with care, the seven-fold increase in post-European sedimentation rates is of the right order to account for the difference between modern mud supply and late Holocene deposition. This infers that most mud delivered to the coast is trapped within the mid-shelf depocentre.

DISCUSSION

In developing a model for modern mud deposition on the Poverty Bay shelf, it is essential to evaluate three key, interrelated elements, namely: sediment supply, the hydraulic regime, and tectonic deformation (Fig. 10).

By New Zealand standards, and indeed global standards, the input of sediment to the Poverty Bay shelf is high. The specific yield of suspended sediment for the region between East Cape and Hawke Bay is 3146 t/km^2 per year, with the Waipaoa River catchment locally contributing 5836 t/km^2 per year (Griffiths 1982; Griffiths & Glasby 1985). These yields are an order of magnitude greater than average yields for all the continents, and rank with other prodigious suppliers of fluvial sediment, namely major islands of the western Pacific rim (Milliman & Meade 1983).

The normal fluvial input to the Poverty Bay shelf, although large, is still less dense than adjacent coastal waters. Consequently, suspended load is dispersed initially in hypopycnal flows. During the course of major floods with return periods of 10 years or more, however, the concentration of suspended sediment may be sufficiently high (>c. 40 kg/m^3) to raise the density of the river water above that of coastal waters, resulting in the formation of hyperpycnal flows. These flows, supplemented by sediment settling from overlying waters, may form fluid mud layers, such as that reported from parts of the Poverty Bay shelf following Cyclone Bola in 1988.

Outside Poverty Bay, on the open shelf, incursions from the south by the Southland Current will periodically force suspended load northwards over the middle shelf, whereas the East Cape Current will invoke a mean southward transport. Superimposed on these broad movements are the effects of local winds. Under southerly winds, sediment in surface waters will be forced northwards along the shelf and obliquely onshore. If these winds are prolonged, downwelling may result, in which case, subsurface return flows may shift suspensates in the lower water column farther offshore. Dispersal under the prevailing northwesterly winds appears to be less important because the middle shelf is partially sheltered by the coast. Nevertheless, southerly dispersal of surface plumes is observed. If northwest winds result in upwelling, any onshore transport of mud at depth is liable to be affected by the shelf anticlines.

Lachlan and Ariel Anticlines, together with Monowai Rocks in the north and Mahia Peninsula in the south, morphologically define the mid-shelf mud belt. In seismic reflection records, the depocentre is depicted as a syncline which follows the general trend of the Mahia Syncline in Hawke Bay (Lewis 1971). From the interpolated subsurface depth of the 18 000 year old wave-cut surface (W1), it is evident that the mid-shelf syncline is depressing differentially. In the south, from off Young Nicks Head to Mahia Peninsula, we estimate a rate of c. 1.5 m per 1000 years, which equates with the subsidence derived for the Mahia Syncline by Lewis (1971). In contrast, the middle shelf north of Poverty Bay may be going down at c. 2 m per 1000 years. This more rapid depression and the likely preferred dispersal of mud to the north account for the enhanced deposition on the northern reach of the Poverty Bay shelf.

A comparison of rates for modern mud supply and late Holocene accumulation argue that most mud is retained on the shelf. Such a scenario concurs with the geological perspective of the shelf (Fig. 10), which depicts the mid-shelf depocentre as a sediment trap with closure in the northeast and southwest by Monowai Rocks and Mahia Peninsula, respectively, and partial closure in the east by Ariel and Lachlan Anticlines. However, an unknown quantity of material suspended in shallow waters may escape over the anticlines, whereas deeper suspensates may pass through the gap separating these folds. Significantly, surficial sediments in the gap are coarser (silty sands) than other middle shelf sediments (Fig. 3), suggesting that a moderate flow may pass through the gap.

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