Source, sea level and circulation effects on the sediment flux to the deep ocean over the past 15 ka off eastern New Zealand

Lionel Cartera,*, Barbara Manighetti a, Mike Elliot a, Noel Trustrumb, Basil Gomez c

a National Institute of Water and Atmosphere, Private Bag 14 901, Kilbirnie, Wellington, New Zealand
b Manaaki Whenua—Landcare Research, Private Bag 11 502, Palmerston North, New Zealand
c Department of Geography and Geology, Indiana State University, Terre Haute, IN 47809, USA

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Abstract

The last post-glacial transgression and present highstand of sea level were accompanied by a reduction in the terrigenous flux to the deep ocean bordering the active convergent margin off the eastern North Island of New Zealand. Although in accord with long-established models of highstand shelf deposition, new data from giant piston core MD97 2121 (2314 m depth) reveal that the flux also varied with terrigenous supply and palaeocirculation. Between 15 and 9.5 ka, the flux reduced from 33 to 20 g/cm²/ka as supply declined with an expanding vegetation cover, and mud depocentres became established on the continental shelf. An increase from 20 to 27 g/cm²/ka during 9.5–3.5 ka coincided with a strengthened East Cape Current which probably introduced sediment from fluvial and shelf sources in the north. The flux profile shows no immediate response to the establishment of modern sea level 7 ka. However, accumulation decreased from 3.5 to 1 ka as more sediments were retained on the shelf, possibly under wind-strengthened, along-shelf currents. Over the last 1 ka, the flux decline halted under increased terrigenous supply during anthropogenic development of the land. Despite the proximity of the North Island’s Central Volcanic Region, major eruptions caused only brief increases (centuries duration) in the terrigenous flux through direct deposition of airfall and possibly fluvial redistribution of onshore volcanic deposits. Frequent earthquakes also had little short-term effect on accumulation although such events, along with volcanism, probably contribute to the long-term high flux of the region. The other measured flux component, biogenic carbonate, reached maxima of 6 g/cm²/ka between 11 and 8.5 ka when nutrient-bearing waters of the East Cape Current dominated the palaeoceanography. After these peaks, carbonate accumulation declined gradually to modern levels of 3 g/cm²/ka.

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1. Introduction

The terrigenous flux to the deep ocean varies with eustatic sea level. The regressive to lowstand phase is usually accompanied by an increase in the terrigenous flux to the continental margin and adjacent ocean basins, whereas the transgressive to highstand phase is a period of reduced flux (e.g., Haq, 1991). Of course, this simple model is modified by other controls that include mechanisms and rates of terrigenous supply to the margin, the size and morphology of the shelf, current and wave...
regimes, tectonism, and volcanism (e.g., Pickering et al., 1989).

Even the eustatic sea level curve may be regionally modified to potentially influence the terrigenous flux. Off New Zealand, for example, the last post-glacial transgression was an episodic event of rapid rises punctuated by stillstands where palaeo-shorelines and seaward thinning sediment wedges developed (Carter et al., 1985, 1986; Carter and Carter, 1986). Some early palaeo-shorelines were occupied for several centuries when substantial wedges > 20 m thick deposited (e.g., Gibb, 1986; Carter and Carter, 1990). These halts in the transgression presumably would have halted or even reversed the decline in offshore supply.

Effects of sea level and other controls on New Zealand oceanic sedimentation are known only in broad terms, there being little differentiation of their relative influences. Within the active convergent margin off the eastern North Island, for example, the ocean received terrigenous sediment during both highstand and lowstand stages of sea level, the principal difference being a reduction in the flux during the transgressive to highstand phase (Carter et al., 2000). However, the roles of other potential controls including tectonism, volcanism, eustasy, stillstands and anthropogenic land development, have yet to be resolved.

Accordingly, this paper is devoted to assessing the changes in the terrigenous flux off eastern North Island over the last 15 ka. This interval was chosen because it encompasses a period of major events that include prominent stillstands in the last post-glacial transgression (Carter et al., 1986), large volcanic eruptions (Carter et al., 1995), a suite of large earthquakes (Berryman, 1993), and human settlement. Another reason for choosing this period is that the flux composition is simple with terrigenous (including volcanic) and biogenic carbonate detritus forming 99% of the core sediment (e.g., Stewart and Neall, 1984). Before 15 ka, the flux also included prominent aeolian and biogenic siliceous components (Nelson et al., 1993; Stewart and Neall, 1984).

2. Sedimentary setting

The study is based on the record preserved in a giant piston core, MD97 2121 which was recovered from a 2314-m-deep slope basin off southern Hawke Bay (Fig. 1). The basin is within the imbricate wedge of the Hikurangi Margin located at the obliquely collisional boundary between the Pacific and Australian plates (Lewis and Pettinga, 1993). Consequently, the continental shelf and slope receive large amounts of terrigenous detritus that is reflected by sedimentation rates of ≥29 cm/ka at Site MD97 2121. Such an influx is attributed to the rapid uplift and exposure of readily erodible rocks onshore, as well as pronounced seismicity and volcanism (e.g., Carter, 1975; Foster and Carter, 1997; Lewis, 1973a,b). A vigorous climate, reflecting the position of the landmass within the Roaring Forties wind belt, has further enhanced sediment erosion and delivery. At present, rivers draining the lower eastern North Island annually carry ~78 Mt of suspended load to the continental shelf (Griffiths and Glasby, 1985).

This combination of tectonically active margin and high sediment input is conducive to large-scale mass failure and turbidity currents (e.g., Barnes and Lewis, 1991; Lewis et al., 1998). However, Site MD97 2121 appears to have escaped the products of redeposition as evinced by (i) an absence of proximal turbidity current channels and mass failure scars as revealed by the swath bathymetric chart of Lewis et al. (1999), (ii) an intact hemipelagic drape as recorded by 3.5-kHz profiles, (iii) a lack of identifiable turbidites and other products of mass failure in the core itself, and (iv) a realistic core stratigraphy (see results).

3. Oceanography

The core comes from beneath a zone of competing surface currents and water masses (inset Fig. 1). The continental shelf is bathed by a northward flowing current (Brodie, 1960) which Chiswell (2000) has

Fig. 1. Location of giant piston core MD97 2121 at the imbricate wedge of the collisional Australian/Pacific plate boundary, with CVR=Central Volcanic Region, PB=Poverty Bay, HB=Hawke Bay, LR=Lachlan Ridge and MS=Mernoo Saddle. Inset shows the main elements of the circulation in particular the Tasman Front/East Australian Current (TF/EAC), East Auckland Current (EAUC), Wairarapa Coastal Current (WCC), East Cape Current (ECC), Wairarapa Eddy (WE), Southland Current (SC), D’Urville Current (DC), Westland Current (WC), Southland Front (SF) and Subtropical Front (STF).
recently termed the Wairarapa Coastal Current (WCC). This flow transports cool, low-salinity water whose character is influenced by (1) Australasian Subantarctic Water, transported to the north of Chatham Rise by the Southland Current (SC), and (2) water from Cook Strait. The volume transport of the WCC is 1.6 Sv off the southern North Island but decreases towards Hawke Bay. There, surface current speeds of 40–50 cm/s have been recorded by Chiswell (2000). East of the WCC is the south-going East Cape Current (ECC) which carries warm, high-salinity subtropical water along the outer continental shelf/upper slope to Chatham Rise where the current turns east along the Rise’s northern flank. Volume transport within the ECC is variable ranging between 10 and 25 Sv. Speeds are similarly variable ranging up to 25 cm/s at 100 m depth, and 10 cm/s at 1000 m (Chiswell and Roemmich, 1998). Seaward of the ECC is the Wairarapa Eddy (Roemmich and Sutton, 1998). This 200-km-diameter, anticyclonic flow is a permanent feature centred on 178°40’ E, 41°00’ S. Its edge, therefore, extends over Site MD97 2121.

Six years of satellite-derived, sea surface temperature data (M. Uddstrom, NIWA; personal communication, 2000.) reveal a marked variability in the opposing currents. The WCC commonly reaches 40°–41°S, but periodically extends to 38°S. Sometimes these major intrusions are accompanied by a seaward displacement of the ECC. Whether such displacements result from the force of the northward intrusion or a weakening of the ECC (Heath, 1985), is uncertain.

4. Methods and data

Core MD97 2121 was taken with a 45-m-long Calypso giant piston corer, deployed from the RV Marion Dufresne during the course of the IMAGES III–IPHIS (Indian and Pacific Ocean Pleistocene and Holocene History) cruise on 30th May, 1997. Once onboard, time constraints permitted only the upper 12 m of the 34.92-m-long core to be analysed for magnetic susceptibility, density (gamma ray counting) and P-wave velocity on a multisensor track (MST) (Nees et al., 1998). Magnetic susceptibility and density measurements were rerun at 2- and 10-cm intervals, respectively, for the whole core at the NIWA laboratory in Wellington. A Bartington MS2 system, with a hand-held probe, measured magnetic susceptibility. Density was determined from a known volume of sediment which was weighed wet, dried at 50 °C and reweighed to obtain the dry weight. Dry bulk density (DBD) was taken as the ratio of dry weight/wet volume, corrected for a salt content of 3.47%. Both data sets correlate well with the shipborne measurements made with the MST system. The upper 5 m of core forms the basis of this study.

Concentrations of calcium carbonate were determined at 10-cm intervals for the whole core using a vacuum gasometric procedure (Jones and Kaiteris, 1983). Each run of 10 subsamples was accompanied by CaCO₃ standards, and the analytical error is calculated to be ±1.0%. The grain size of subsamples from 10-cm intervals was measured with a Sedigraph 5100 particle size analyser with a precision of ±0.1 μm (e.g., Robinson and McCave, 1994).

To detect any anthropogenic influence, the upper 1.1 m of the core was scrutinised for pollen using techniques outlined in Moore et al. (1991). Sampling intervals were 1 cm for the top 0.5 m and 3 cm for the remaining 0.6 m. The resultant pollen profiles were used as indicators of altered land use under Polynesian and later European settlement of New Zealand.

A dual approach was taken to establishing a core stratigraphy. Tests of the planktonic foraminifer, Globigerina inflata, were hand-picked from eight, 1-cm-thick slices and radiocarbon dated (Table 1; Fig. 2). These data are supplemented by a tephrochronology derived from tephra layers which were correlated with dated counterparts onshore using glass shard chemistry and heavy mineralogy (e.g., Froggatt and Lowe, 1990; Carter et al., 1995). A linear regression for the radiocarbon/tephra dates against core depth exhibited a high correlation with R = 0.998 (Fig. 2). All ages in this paper are expressed in calendar years according to the Calib. v.4.2 programme of Stuiver and Reimer (1993) with the marine calibration data set of Stuiver et al. (1998) and a 30-year moving average to account for possible age variance of planktonic foraminifera within a 1-cm-thick core subsample. Tephra dates were calibrated using the atmospheric decadal calibration data set of Stuiver et al. (1998) and a Southern Hemispheric correction of −24 years.

Compositional, physical and chronostratigraphic data allow estimation of the sedimentary burial flux.
or mass accumulation rate (MAR) according to the relationship:

\[
\text{MAR} \left( \frac{g}{cm^2/ka} \right) = \left( \text{dry bulk density g/cm}^3 \right) \times \left( \text{linear sedimentation rate cm/ka} \right) \times \left( \frac{\text{grams of component}}{\text{grams dry bulk sediment}} \right)
\]

Linear sedimentation rates are based on the radiocarbon and tephra chronologies. These rates were not corrected for deposition of volcanic airfall. This effect cannot be quantified with confidence because:

1. Even though site MD97 2121 is only 200 km from the Central Volcanic Region, no well-defined macroscopic tephra was found in the 15-ka-old section of the core. This paucity reflects the site’s position just outside the main eastward dispersal path for erupted ash (Carter et al., 1995). Furthermore, what ash accumulated, was mixed into the sediment because the tephra was too thin (\(< 1 cm\) thick; Kennett, 1981) to smother the bioturbating benthos.
2. Following an eruption, rivers draining ash-covered landscapes delivered volcanic-rich loads to the shelf and beyond (e.g., Lewis and Kohn, 1973; Pillans et al., 1993). Once deposited, such loads would be difficult to distinguish from bioturbated airfall.

Table 1
AMS radiocarbon and tephra ages in radiocarbon and calendar years

<table>
<thead>
<tr>
<th>Tephra</th>
<th>Depth (cm)</th>
<th>Tephra</th>
<th>(^{14}C) age (year)</th>
<th>(\pm)</th>
<th>Cal. age (years BP)</th>
<th>Source of (^{14}C) age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Taupo</td>
<td>54–55</td>
<td>1800</td>
<td>15</td>
<td>1718</td>
<td>Sparks et al., 1995</td>
<td></td>
</tr>
<tr>
<td>Waimihia</td>
<td>106–107</td>
<td>3280</td>
<td>20</td>
<td>3472</td>
<td>Froggatt and Lowe, 1990</td>
<td></td>
</tr>
<tr>
<td>Whakatane</td>
<td>190–191</td>
<td>4830</td>
<td>20</td>
<td>5580</td>
<td>Froggatt and Lowe, 1990</td>
<td></td>
</tr>
<tr>
<td>Tuhua</td>
<td>270–271</td>
<td>6130</td>
<td>30</td>
<td>6970</td>
<td>Froggatt and Lowe, 1990</td>
<td></td>
</tr>
<tr>
<td>Rotoma</td>
<td>342–343</td>
<td>8530</td>
<td>10</td>
<td>9520</td>
<td>Froggatt and Lowe, 1990</td>
<td></td>
</tr>
<tr>
<td>Poronui</td>
<td>408–409</td>
<td>9810</td>
<td>50</td>
<td>11190</td>
<td>Froggatt and Lowe, 1990</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Radiocarbon</th>
<th>Depth (cm)</th>
<th>NZA Ref. number</th>
<th>(^{14}C) age (year)</th>
<th>(\pm)</th>
<th>Cal. age (years BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>35–36</td>
<td>10069</td>
<td>1139</td>
<td>56</td>
<td>672</td>
<td></td>
</tr>
<tr>
<td>40–41</td>
<td>9872</td>
<td>1748</td>
<td>56</td>
<td>1287</td>
<td></td>
</tr>
<tr>
<td>58–59</td>
<td>9871</td>
<td>2203</td>
<td>56</td>
<td>1808</td>
<td></td>
</tr>
<tr>
<td>110–111</td>
<td>9859</td>
<td>3241</td>
<td>56</td>
<td>3058</td>
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</tr>
<tr>
<td>196–197</td>
<td>9858</td>
<td>5379</td>
<td>57</td>
<td>5729</td>
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</tr>
<tr>
<td>310–311</td>
<td>10070</td>
<td>8286</td>
<td>61</td>
<td>8828</td>
<td></td>
</tr>
<tr>
<td>420–421</td>
<td>10071</td>
<td>10818</td>
<td>70</td>
<td>12232</td>
<td></td>
</tr>
<tr>
<td>480–481</td>
<td>9857</td>
<td>12265</td>
<td>96</td>
<td>13816</td>
<td></td>
</tr>
</tbody>
</table>

Fig. 2. Age–depth curve based on dated tephra and radiocarbon ages, all of which are expressed in calendar years determined from Calib. v.4.2 (Stuiver and Reimer, 1993; Stuiver et al., 1998). Tephra are denoted as Po=Poronui, Rt=Rotuma, Tu=Tuhua, Wk=Whakatane, Wm=Waimihia and Tp=Taupo. Calibrated and radiocarbon ages are presented in Table 1.
5. Chronostratigraphy

Using the magnetic susceptibility profile as a guide, sections of core with prominent magnetic peaks were subsampled for glass shards and/or heavy minerals. The following tephras were identified (Table 1); Poronui (11.19 ka), Rotoma (9.52 ka), Tuhua (6.97 ka), Whakatane (5.58 ka), Waimihia (3.47 ka) and Taupo (1.718 ka). With two exceptions, the AMS radiocarbon ages from *G. inflata* agreed well with the tephra chronology (Fig. 2) to provide good time control (Figs. 3–5). The two exceptions may reflect uncertainties in the surface water $^{14}$C reservoir age as recently highlighted by Sikes et al. (2000).

Using the intervals defined by the chronostratigraphy, mean linear sedimentation rates were calculated to range from 29 to 39 cm/ka. Rates increase irregularly upcore from 36 to 39 cm/ka between 15 and 3.5 ka. Thereafter, rates decreased to a modern level of 31 cm/ka.

6. Physical and chemical properties

6.1. Core description

In hand specimen, core MD97 2121 is comprised mainly of light olive grey (5y 5/2) hemipelagic mud. Bedding and laminae are not obvious; a feature that is consistent with the bioturbatory mottling. Concentrations of black, pyritic-like flecks are common. The upper 5 m of core has poorly preserved tephras that either are reduced to ill-defined blebs or thoroughly bioturbated into the sediment to produce an ash-bearing mud that is identifiable only by microscopy, bulk density or magnetic susceptibility. Below 5 m core depth, tephras are better preserved as distinct silty to sandy laminae.

6.2. Magnetic susceptibility

Profiles of magnetic susceptibility clearly identify tephra-bearing sediments which appear as distinct

![Fig. 3. Magnetic susceptibility, dry bulk density and percentage calcium carbonate for MD97 2121. Of note are the prominent but brief disturbances associated with major rhyolitic eruptions annotated as in Fig. 2.](image)
peaks (Fig. 3). The Whakatane Tephra produces the largest spike whereas the Rotoma and Taupo tephras have a subdued signal that is barely above the background level.

6.3. Dry bulk density

Dry bulk density (DBD) decreases upcore from 0.92 g/cm³ at the base of the 15-ka section to 0.68 g/cm³ in the Recent. The profile is broken by peaks corresponding to dispersed tephra from the Rotoma, Whakatane and Waimihia eruptions. The presence of dense glass and heavy minerals elevates DBDs by up to 0.06 g/cm³ above background levels. By comparison, the Taupo eruptives barely disrupt the DBD profile because of low ash concentrations. Nevertheless, airfall and waterborne volcanic detritus increase DBDs and may dominate the effect of calcium carbonate on sediment density. In oceanic sediments, DBDs commonly correlate with carbonate contents, exhibiting correlation coefficients of \( R=0.84–0.96 \) (e.g., Lyle and Dymond, 1976; Snoeckx and Rea, 1994). However, volcanic-bearing sediments off eastern New Zealand have a less robust DBD/CaCO\(_3\) correlation of \( R=0.71 \) (Carter et al., 2000). For MD97 2121, the correlation is weak with \( R<0.4 \).

6.4. Calcium carbonate

Carbonate contents are low with a mean \( x=13.4\% \) (\( \sigma=2.81; n=57 \)). Contents generally fluctuate within a range of 11–19% apart from localised reductions to <11%, which result from dilution by the Whakatane, Waimihia and Taupo tephras (Fig. 3). Overall, carbonate increases upcore to a maximum of 25% near 9 ka. Thereafter, it reduces to \( \sim 12\% \) at 6.5 ka and hovers around that concentration until modern times when contents rise slightly to 14%.

7. Textural properties

The prevailing sediment is silty clay with an average composition of sand 4.6%, silt 33.7% and...
clay 61.6% (Fig. 4). From a broad perspective, there is little variability in the gross texture. Sand, for example, is restricted to between 2% and 5%, except for spikes related to volcanic ash. The Whakatane airfall, in particular, caused the sand content to rise to 27%. By comparison, the Taupo eruption had little impact on the coarse texture presumably because so little sand-sized ash reached the site (Carter et al., 1995).

When examined in detail, some textural parameters exhibit a series of subtle upcore changes. Apart from coarse modes associated with dispersed tephra, the dominant mode reduces irregularly upcore from 4.4 μm at 13.5 ka to 2.5 μm at 9.7 ka. From then on, the mode increases to 4 μm at ~ 7 ka, and oscillates about the silt/clay boundary into the present. With regard to silt, this component increases slightly upcore from 29% to 31% between 12 and 7 ka, before increasing more rapidly to 37% at the core top. These changes are mirrored in the clay profile with contents decreasing from 65% at 12 ka to 60% in the present. Superimposed on these broad trends are smaller shifts in texture.

8. Flux rates

The terrigenous flux on the Hikurangi Margin has remained high but variable (Fig. 5). About 13.5 ka—the boundary between marine isotope stages (MIS) 1 and 2—the flux was ~ 32 g/cm²/ka, declining to ~ 18 g/cm²/ka at 8.7 ka. Later, the trend reversed with a rate of 26 g/cm²/ka at 3.5 ka, only to reduce to 18 g/cm²/ka at 1 ka and remain there into Recent times. Superimposed on this profile is a series of positive peaks corresponding to the addition of airfall (and waterborne?) ash to the oceanic flux. As this material has been bioturbated into the sediment, the original amount of the airfall is uncertain, and hence has been incorporated into the flux. However, judging by the thinness (typically ~ 1 cm) of preserved macrotephra further down in the core, inclusion of the dispersed tephra will not substantially alter linear sedimentation rates and hence MARs.

Compared to the terrigenous flux profile, the biogenic carbonate counterpart is more irregular (Fig. 5). Localised reductions in carbonate flux, coincident with concentrations of volcanic ash, are interspersed with
Fig. 6. Pollen frequency spectra for selected taxa for the last 3.5 ka highlighting the strong bracken (*Pteridium*) signal associated with anthropogenic influences. Ages from radiocarbon and the Taupo (Tp) and Waimihia (Wm) tephras.
increments. Nevertheless, the flux profile displays a general upcore trend from 4.5 g/cm²/ka in early MIS 2 to maxima of 6 g/cm²/ka at between 11 and 8.5 ka. Later, the profiles show an irregular decline to 3 g/cm²/ka near the core top.

9. Palynology

The palynological record for the last 3.5 ka comprises three distinct pollen zones identified from cluster analysis (Fig. 6). The basal zone HB3 (3.5–1.4 ka) shows that the vegetation of the Hawke’s Bay region was dominated by podocarp–beech forest with *Dacrydium cupressium*, *Prumnopitys taxifolia* and *Nothofagus* spp. as the main species (Wilmshurst et al., 1997). In 1.718 ka, the forest cover was damaged by ignimbrite and ash from the Taupo eruption—an event that was followed by a flourish of the bracken, *Pteridium esculentum*. However, forest recovery was swift and the cover was complete within 200 years as also confirmed by Wilmshurst and McGlone (1996).

Pollen zone HB2 (1.4 to ~0.3 ka) is distinguished by a flood of bracken pollen that is more pronounced than the increase following the Taupo eruption—an event that was followed by a flourish of the bracken, *Pteridium esculentum*. However, forest recovery was swift and the cover was complete within 200 years as also confirmed by Wilmshurst and McGlone (1996).

Zone HB1 (~0.3 ka–present) has, in addition to its high bracken content, strong representation from herbs and climbers which have increased at the expense of podocarp–beech forest species. HB1 is further distinguished by the first presence of exotic pollen including *Pinus radiata* which was planted extensively in Hawke’s Bay in the 1940s (Guthrie-Smith, 1969).

10. Discussion—variability and controls over the ocean flux

10.1. Terrigenous flux

Three bouts of accumulation are evident; Phase I (~15–9.5 ka), Phase II (~9.5–3.5 ka) and Phase III (~3.5 ka–present) (Fig. 7; note that a 17.6-ka-long record is presented in the figure to outline the flux trend leading up to 15 ka). Superimposed on these broad phases are short-term fluctuations, some of which are related to the direct accumulation of volcanic airfall (Figs. 5 and 7). Volcanism has the potential to further affect the flux through post-eruption erosion of terrestrial airfall and non-welded ignimbrites (e.g., Pillans et al., 1993; Wilson and Walker, 1985). To evaluate any longer term response, the total volume of onshore and offshore volcanic airfall was compared to the flux trends of Phases I–III (Fig. 7). According to the data of Carter et al. (1995) and Froggatt and Lowe (1990), approximately 48 km³ of airfall ash was erupted during Phase I, but this failed to halt a decline in the flux (Fig. 7). This lack of response is confirmed in the later phases. The terrigenous flux actually increased in Phase II even though the airfall reduced to 31 km³, and decreased in Phase III when the airfall peaked at 81 km³. Phase III is particularly noteworthy because it encompasses the largest eruption in MIS 1—the Taupo event. About ~50 km³ of airfall and 70 km³ of ignimbrite were expelled causing widespread destruction of the forest with subsequent increase in erosion and river input to the shelf (Wilson and Walker, 1985; Wilmshurst and McGlone, 1996; Wilmshurst et al., 1999).

Yet, the eruption had little impact on the oceanic flux. Three factors appeared to have buffered any impact. First, MD97 2121 is located just outside the main dispersal path of Taupo airfall as mapped by Carter et al. (1995) and Wilmshurst and McGlone (1996). Second, despite widespread destruction of the forest, the palynological record (Fig. 6 of this study; also Wilmshurst and McGlone, 1996) shows rapid regeneration of forest, with complete restoration within ~200 years of the eruption. Third, the Taupo event was well into the modern highstand, so that river-transported volcanic detritus is more likely to be retained on the shelf.

Earthquakes are also a potential cause of short-term fluctuations in the flux through destabilisation of hill slopes and alteration of river base levels. Destabilisation and mass failure of slopes in river catchments may temporarily increase the fluvial input to the coast (e.g., Goff, 1997). Evidence of past earthquakes in the region is provided by suites of tectonically uplifted coastal terraces which yield a record of major palaeoseismic events back to 6.7
ka—the limit of the record (Fig. 7; Berryman, 1993; Berryman et al., 1989). Uplifted terraces between Poverty Bay and southern Hawke Bay indicate at least five events between 6.7 and 3.5 ka in Phase II, and five events in Phase III. There is little difference in the frequency or magnitude of events in Phases II and III even though these were periods of contrasting flux. Furthermore, individual seismic events failed to produce discernible pulses in the flux profile (Fig. 7). A similar lack of sedimentary response was noted in limnological records from Hawke’s Bay (Eden and Page, 1998).

While earthquakes (and volcanism) probably help maintain the overall high rate of terrigenous sedimentation at the North Island imbricate wedge, they are not controlling Phases I–III. Other factors, including the last post-glacial transgression, margin morphology, regional palaeoceanography, terrestrial vegetation, and climate can play a role as outlined in the following discussion.

### 10.2. Phase I (~15–9.5 ka)

This phase has a declining terrigenous flux that is part of a longer term trend beginning around 17.6 ka after a period of high accumulation during the last glacial maximum (Carter et al., 2000; Stewart and Neall, 1984). The decline coincided with a period of rapid environmental change when several factors conspired to reduce the terrigenous input to the ocean.

1. Rising sea level progressively entrapped sediment on the upper continental margin as shown by Lewis (1973a). His isopach charts clearly define a landward migration of depocentres over the southern Hawke Bay upper slope and shelf during the last post-glacial transgression. Pauses in the transgression, such as the 13.5- and 12.5-ka stillstands at -56 and -46 m depth, respectively (Fig. 7), failed to significantly affect the oceanic flux, although subtle effects could be masked by volcanic influences on the flux as is the...
case for the \( -56 \text{-m} \) event which coincides with accumulation of an unidentified tephra (Fig. 7). In general terms, stillstands could be expected to halt or even reverse the decline in the oceanic terrigenous flux as sediment escaped beyond the palaeoshoreline. However, rates of escape depend upon the position of the palaeoshoreline relative to the shelf edge. In Hawke Bay, the \( -56 \text{-m} \) palaeoshoreline was \( \sim 45 \text{ km} \) from the shelf edge, and was positioned landward of the prominent growing anticline of Lachlan Ridge that extends across the bay entrance, near the outer shelf (Lewis, 1971, 1973a). Thus, sediment from local Hawke’s Bay rivers—the closest major point sources to MD97 2121—was largely retained on the shelf.

(2) The palaeocirculation during Phase I comprised two competing current systems (Nelson et al., 2000; Weaver et al., 1998). Foraminiferal assemblages and \( \delta^{13}C \) signatures of surface waters indicate northward incursions of Australasian Subantarctic Water, carried by the ancestral Southland Current (SC), to be competing with south-flowing subtropical waters transported by the ECC. At that time, the ECC was probably weaker than now because the main driving force behind the flow, namely the Tasman Front/East Australian Current (Stanton et al., 1997) was located well north of its present position at \( \sim 32^\circ \text{S} \). During the last glacial maximum, the front/current was positioned at \( \sim 26^\circ \text{S} \) (Martinez, 1994). At that latitude, its entry to New Zealand waters was probably inhibited by Norfolk Ridge. Any diversion of the Tasman Front/ East Australian Current away from New Zealand would cause a downstream weakening of the ECC thereby facilitating its off-shelf displacement by northward incursions of subantarctic water (e.g., Roemmich and Sutton, 1998). These incursions were also favoured by an intensified glacial wind regime (e.g., Stewart and Neall, 1984) that would have forced the ancestral SC in a fashion similar to its modern counterpart (Carter and Herzer, 1979). Furthermore, lowered sea level would have displaced the along-shelf SC towards the Southland Front, positioned along the shelf edge and upper slope (Chiswell, 1996). Frontal currents would have been reinforced by the SC and by the pronounced temperature gradients of the MIS 2 ocean (Chiswell, 1996; Weaver et al., 1998). Any modification of the SC by water from Cook Strait to form an ancestral Wairarapa Coastal Current is unlikely. Cook Strait had only just opened and a full oceanic flow through the Strait had not yet established (Proctor and Carter, 1989).

In terms of terrigenous loads, a weakened and seaward-displaced ECC transported less sediment to MD97 2121 from the major rivers north of Hawke Bay. In addition, any load carried by the ECC could be diluted by SC-transported subantarctic water. This water mass was likely to carry less sediment than the ECC by virtue of the distal location and smaller loads discharged by South Island rivers feeding the SC, compared to more proximal and larger rivers feeding the ECC. Towards the end of Phase I, the influence of the subantarctic SC waned and the subtropical ECC prevailed as evinced by an increase in warm water transitional foraminifera at the expense of subpolar to polar species (Weaver et al., 1998).

(3) Changes in the terrestrial vegetation towards a more permanent cover, under a rapidly warming and less windy climate (McGlone, 2001; McGlone et al., 1994), probably reduced erosion. For the last glacial maximum, Pillans et al. (1993) noted that the prevailing grassland and scrub communities were frequently disturbed by climatic extremes and fire—conditions that encouraged erosion. In contrast, the period \( \sim 15 – 11.5 \text{ ka} \) witnessed a rapid expansion of podocarp/ hardwood forests to cover roughly 90% of the land below the tree line (McGlone et al., 1994).

(4) Finally, the aeolian input decreased about 11 ka when the offshore flux of aerosolic and loessic quartz went from \( \sim 0.7 \) to \( \sim 0.3 \text{ g/cm}^2/\text{ka} \) (data of Fenner et al., 1992; Stewart and Neall, 1984). However, the impact of this reduction was small as the aeolian component was <3% of the terrigenous flux at that time.

10.3. Phase II (\( \sim 9.5 – 3.5 \text{ ka} \))

From \( \sim 9 \text{ ka} \), the terrigenous flux gradually increased up-core from \( \sim 18 \) to 26 \( \text{g/cm}^2/\text{ka} \) at 3.5 ka (excluding the localised effects of the Whakatane and Waimihia tephras; Fig. 7). This reversal happened despite a rising sea level and the shoreward migration of shelf depocentres (Lewis, 1973a). Furthermore, the increase continued unabated showing no immediate response to the establishment of modern sea level about 7 ka (Gibb, 1986). This lack of a discernible response suggests other influences on the ocean flux.
Increased Phase II accumulation began soon after inception of a stronger ECC. Elevation of sea surface temperatures from 10 to 14 °C between 13.5 and 11 ka (Nelson et al., 2000), a distinct change to a subtropical δ13C signature in *G. bulloides* about 11 ka (Nelson et al., 2000), and a strengthening of the warm-water *transitional* foraminiferal assemblage (Weaver et al., 1998), collectively point to re-establishment of the full subtropical inflow by ~11–10 ka (Fig. 7). A stronger circulation may be reflected by a coarsening of the dominant sediment mode at that time (Fig. 4). Resurgence of the ECC was probably fuelled by the return of the Tasman Front/East Australian Current to its modern latitude at ~32°S (Martinez, 1994). Relocation allowed this current system to pass through a major gap in the Norfolk Ridge thereby rejuvenating the ECC and adjacent Wairarapa Eddy (inset Fig. 1). We argue that a stronger ECC would increase delivery of terrigenous sediment from the shelf and upper slope, north of MD97 2121, where rivers have some of the highest discharge rates in New Zealand (Griffiths and Glasby, 1985). The current would carry sediment injected directly from point sources plus detritus reworked from the seabed as manifested by the presence of current-scoured moats along the seaward base of Lachlan Ridge off Hawke Bay (Barnes et al., in press).

The prevailing westerly wind regime was weaker than now, and appears to have borne a strong northerly component (McGlone et al., 1994; Shulmeister, 2001). Under such conditions, the curl of the wind stress over the shelf probably directed wind-driven currents offshore, possibly assisting oceanic dispersal of terrigenous detritus.

Changes in the vegetation cover were not a factor behind the increased flux. Early in Phase II, reforestation of lowland areas was complete under a climate that was warmer than now (McGlone et al., 1994). Apart from short-term disturbances by volcanic eruptions (Wilmshurst et al., 1999), the forest cover was permanent. Such warm conditions are consistent with the proposed strengthened ECC and Wairarapa Eddy.

10.4. Phase III (3.5 ka–present)

After peaking at ~26 g/cm²/ka about 3.5 ka, the flux declined steadily to ~18 g/cm²/ka at 1 ka, and then maintained that rate to the present. The decline occurred approximately 3000 years after sea level had achieved its modern position. Either there was a substantial time-lag before highstand entrapment affected the flux or there was a decline in sediment supply to the shelf. Any decline was unlikely to be related to a vegetation change as the forest cover had already reached its maximum extent between 9 and 10 ka (McGlone, 2001). Again, forests suffered short-term disturbances during volcanic eruptions (Wilmshurst et al., 1999) but were otherwise permanent until human settlement of the region.

We suggest that the reduced flux was affected by climate-forced changes in the shelf circulation. Following the warm period ~9 ka, the climate cooled to modern levels. McGlone et al. (1994) and Shulmeister (2001) speculate that reduced temperatures resulted from an increased southerly component of the strengthening regional westerly wind flow. Under those conditions, the Ekman drift would help constrain sediment to inner shelf depocentres, thus depriving the deep ocean (e.g., Carter and Herzer, 1979; Foster and Carter, 1997).

Whether or not changes in the circulation were associated with the stabilised terrigenous flux from 1 ka to present, is equivocal. Certainly, stabilisation coincides with the onset of human settlement, as shown in the pollen record by a sharp decline of podocarp/hardwood forest species concomitant with a major and prolonged increase in bracken (Fig. 6; McGlone, 1989; Wilmshurst, 1997). Initially, bracken was used as a food source by Polynesian settlers who first settled between 500 and 1000 years BP (McGlone, 1989), although the MD97 2121 data suggest settlement in Hawke’s Bay may have been earlier by ~400 years. Forest clearance and soil erosion continued with the European settlement of the region in the late 19th century as attested by contemporaneous increases in sedimentation within lakes and coastal embayments (e.g., Goff, 1997; Page et al., 1994; Wilmshurst, 1997).

Similar anthropogenic-timed increases in sedimentation occurred on the nearby Poverty Bay shelf (Wilmshurst et al., 1999) and off southern Hawke Bay (Barnes et al., 1991). It is unlikely that the higher terrigenous flux is caused by exceeding the trapping capacity of the nearshore sediment prism because (1) this prism is sand dominated (Lewis, 1973b; Pantin, 1966) and (2) hypopycnal and hyperpycnal flows at
major point sources appear to bypass the prism in favour of middle shelf and more seaward depocentres (Foster and Carter, 1997).

Despite the present highstand, substantial quantities of sediment still escape the shelf. Presently, \( \sim 18 \text{ g/cm}^2/\text{ka} \) leaves the shelf for MD97 2121, this rate being approximately half that of the last glacial lowstand (e.g., Carter et al., 2000). By comparison, the South Island highstand shelf retains much of its fluvial and coastal input to the extent that the nearby Bounty Trough receives mainly calcareous biopelagic sediment (Carter and Carter, 1993). This retention of terrigenous material reflects the shelf width which averages \( \sim 60 \text{ km} \) south of Chatham Rise, and the vigorous along-shelf circulation (Carter and Herzer, 1979). Furthermore, the terrigenous input to the South Island shelf reduced rapidly \( \sim 11 \text{ ka} \) or earlier with the deglaciation of large lakes and their subsequent transformation into effective fluvial sediment traps (Carter and Carter, 1990). On the other hand, the eastern shelf off the lower North Island is narrower, averaging 30 km width, and is subject to a less vigorous circulation due to its position at the northern and less extreme sector of the Roaring Forties wind belt. In addition, this shelf has a higher fluvial input (\( \sim 78 \text{ Mt/a} \) of suspended load—data of Griffiths and Glasby, 1985) than the South Island, south of Chatham Rise (17 Mt/a).

10.5. Biogenic carbonate accumulation

From MIS 2 to MIS 1 at MD97 2121 and other sites east of the North Island (Carter et al., 2000; Fenner et al., 1992; Stewart and Neall, 1984), the biogenic flux was highest towards the end of the glacial period and then declined into the following interglacial. This peak is a reflection of glacial carbonate productivity associated with the injection of nutrient-rich waters from southern subantarctic sources, coupled with local wind-induced upwelling off the lower North Island; all these effects occurring under the vigorous climatic conditions of MIS 2 (Nelson et al., 2000; Weaver et al., 1998).

In contrast, the carbonate flux declined irregularly through MIS 1 after reaching maxima between 11 and 8.5 ka (Fig. 5). Unlike the maximum in MIS 2, these younger events coincided with increasing warmth as the subtropical ECC and Wairarapa Eddy began to dominate the circulation. Both these flows are likely to improve local nutrient levels and production; the ECC appears to transport upwelled water from East Cape (Bradford and Chapman, 1988), and the warm-core Wairarapa Eddy may be a zone of elevated plankton productivity (Bradford et al., 1982; Chiswell and Roemmich, 1998). Following the Holocene warm period, the vegetation history suggests climate cooling (McGlone et al., 1994) which may be equated with a less intense subtropical circulation and, by association, reduced productivity.

11. Conclusions

Long-standing models of highstand/lowstand sedimentation call for a reduction in the terrigenous flux to the deep ocean during transgressive/highstand conditions as sediment becomes progressively retained within coastal and shelf depocentres. While supporting this general trend, profiles of terrigenous flux from MD97 2121 show other effects that relate to changes in the palaeocirculation and sediment supply.

1. From \( \sim 15–9.5 \text{ ka} \), the terrigenous flux in the deep ocean declined as sea level rose and a permanent vegetation cover established onshore.

2. The flux increased from \( \sim 9.5–3.5 \text{ ka} \) as the subtropical ECC strengthened and delivered sediment from northern shelf and fluvial sources. Surprisingly, establishment of modern sea level was not immediately recorded in the flux profile suggesting that supply and dispersal initially outweighed stillstand trapping effects.

3. A reduction in the flux from 3.5 to 1 ka may reflect a waning ECC, but a subsequent stabilisation of the flux from 1 ka to present may be a response to greater erosion accompanying the human colonisation of New Zealand. Despite entrapment on the shelf and within slope basins, a substantial amount of sediment still escapes the shelf on account of its narrowness and weak circulation.

4. Superimposed on these changes were effects of volcanic eruptions which, although frequent and large, caused only short-term fluctuations (centuries duration) in the flux.

5. Major earthquakes are known from uplifted coastal terraces but they have no obvious immediate effect on the flux. But in the long term, tectonism and volcanism contribute to the high terrigenous input.
(6) The biogenic flux initially increased into MIS 1 to maxima ~11–8.5 ka which is consistent with inception of a strong subtropical inflow. Production may have been encouraged by either nutrient-rich waters carried by the ECC from zones of upwelling, or by the influence of the proximal warm core Wairarapa Eddy, or both these factors.

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