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Late Quaternary geomorphology of the Manawatu coastal plain, North Island, New Zealand

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A R T I C L E I N F O

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A B S T R A C T

This review examines the geomorphological evolution of the Manawatu coastal plain from the Last Glacial Maximum through the Holocene post-glacial marine transgression (c. 7500 cal yr BP) until the present, providing a context within which to examine the extent to which the Holocene evolution of the lower Manawatu valley follows tripartite evolutionary models of incised-valley infill. During the last glaciation the Manawatu River incised a broad, deep valley, while tributaries within the catchment dissected relict marine terraces. This occurred concurrently with the formation of a parabolic dune field at Koputaroa, with sediment sourced from the bars of the Manawatu River. During the post-glacial marine transgression rising sea-levels inundated the incised-valley. At the culmination of the Holocene marine transgression, the Himatangi anticline and Poroutawhao High restricted oceanic influences in the proto-Manawatu estuary, which was partially infilled with estuarine sediment. Estuarine deposition was quickly succeeded by fluvial deposition and floodplain due to high volumes of fluvial sediment introduced by the Manawatu River. A lack of characteristic bay-head delta development may be explained by rapid infill and consequent sediment bypass.

1. Introduction

The Manawatu region lies on the southwestern coast of the North Island, New Zealand, within the c. 40,000 km^2 Plio-Pleistocene Wanganui Basin (Figs. 1 and 2). The geomorphology of the region is dominated by incised-valleys, estuaries, a prograding coastal plain and transgressive dune field, and aggrading floodplains, all influenced by sea-level fluctuations over the last full glacial cycle (Fig. 3). Today, the Manawatu River estuary covers an area of approximately 2 km^2. The estuary itself comprises a sandy spit with well developed parabolic dune fields. Impounded behind the spit are extensive estuarine sand- and mudflats.

This review examines published research that has investigated the geomorphological evolution of the Manawatu coastal plain, through the Last Glacial Maximum (LGM) and Holocene, to provide a context within which to examine the extent to which the Holocene evolution of the lower Manawatu valley follows tripartite evolutionary models of incised-valley infill. This period is marked by dramatic climatically induced eustatic sea-level changes, which has acted as a major driver on geomorphic processes in the Manawatu. During the LGM lowered sea-levels resulted in fluvial incision into Plio-Pleistocene sediments throughout the region. Sedimentary infill of the incised-valleys occurred as rising sea-levels during the Holocene post-glacial marine transgression transported sediment landward in conjunction with significant infilling of accommodation space by fluvial processes. Following the culmination of the marine transgression sedimentary processes at the coastal margin were dominated by a landward progradation of the most extensive transgressive dune field in New Zealand, fed by transgressive sand sheets migrating with rising sea-levels.

2. Regional and geological setting

The Manawatu region is subject to a temperate maritime climate, with rainfall increasing from 800 mm at the coast, to more than 5000 mm along the inland ranges. There is no marked seasonality (Heerdegen and Shepherd, 1992). At the coast, waves approach the Manawatu from either the southwest or west. From the southwest, waves generated in the northern Cook Strait have a maximum fetch of 100 km, resulting in moderate wave energy. Waves approaching from the west (Tasman Sea) have a much greater fetch, resulting in a dominant westerly swell. North of the Wanganui River the westerly swell approaches the coast at an oblique angle resulting in a southward moving long-shore drift.

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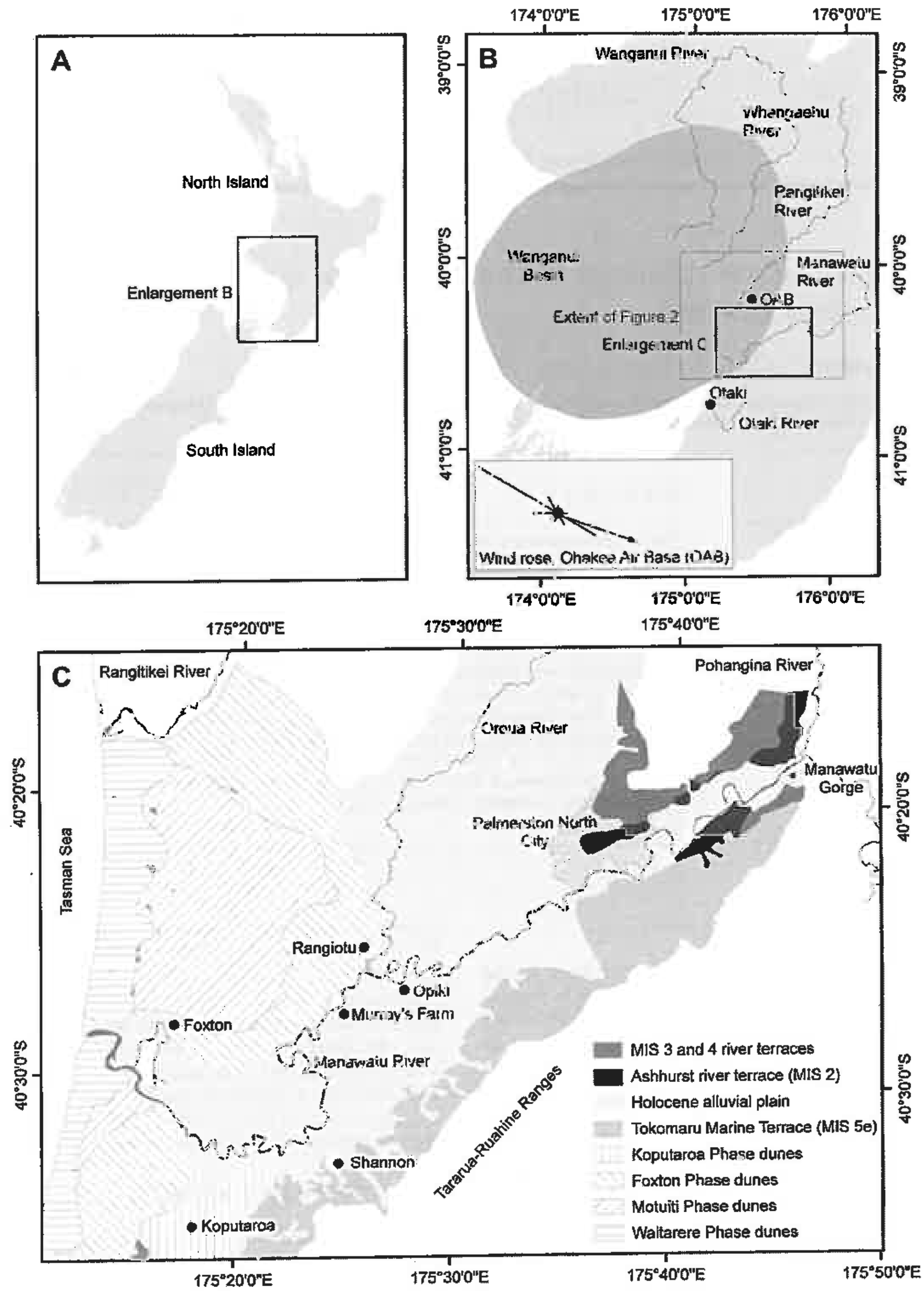

Fig. 1. (A) Contextual map of New Zealand; (B) location map of the Wanganui Basin, regionally significant rivers and localities and location of Fig. 2; (C) quaternary geology of the Manawatu region (after Shepherd, 1987).

Accordingly, waves in the nearshore not only transport large quantities of sand-sized sediment shoreward from the northwest, but southwards as well due to long-shore drift.

North of the Wanganui River coastal sediments are dominated by volcanic detritus (hornblende, hypersthene and augite) sourced from the Taranaki and Central Volcanic regions in the north, transported to the coast by rivers and carried southwards by dominant littoral drift (Gibb, 1977; Muckersie and Shepherd, 1995). South of the Wanganui River the present coastline is dominated by fine-grained sandy sediment derived from Plio-Pleistocene Wanganui Basin sediments and the greywacke axial ranges to the east. Here, the coast is characterised by low gradients due to the fine-

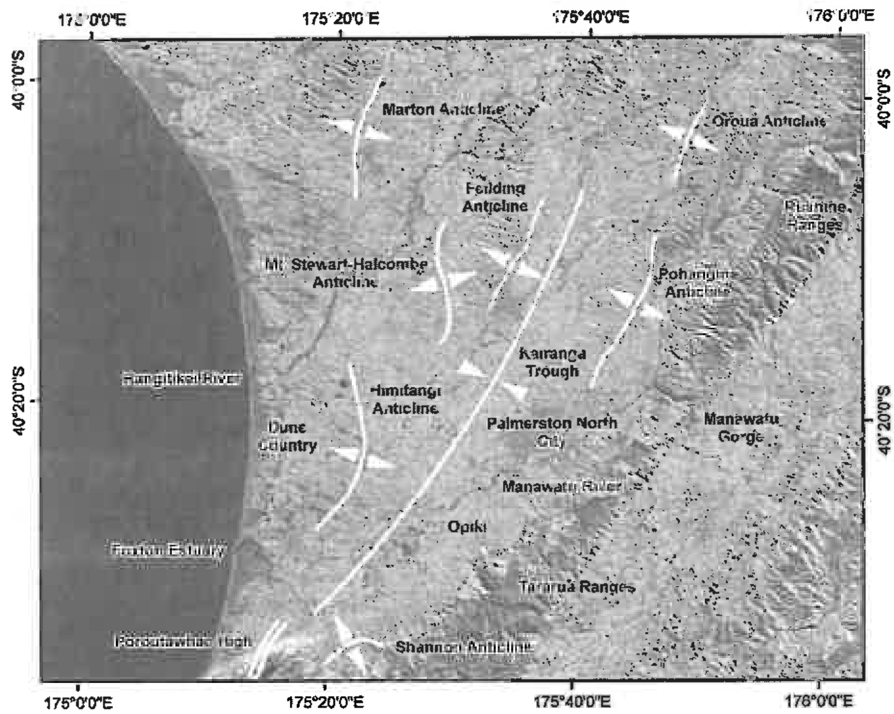

Fig. 2. Fold axes of the Manawatu anticlines overlaid on an Enhanced Thematic Mapper image of the Manawatu region (after Clement and Brook, 2008).

grained nature of the sediment, a relatively low tidal range and moderate wave energy (Shand et al., 2001). The beaches themselves tend to have modally intermediate (ripped) to dissipative (uniform) configurations with lower cross-shore slopes, and are typically backed by a foredune which in places reach several metres in height. The morphology of the foredune varies alongshore from well vegetated stabilised incipient and established foredunes to poorly vegetated blown-out and hummocky (Hesp, 2001). In stark comparison beaches south of the Otaki River are composed of a mix of fine-grained sand and fluvially derived gravels, resulting in narrower, steeper and more reflective beaches, usually featuring a developed gravel berm and less developed foredunes.

The eastern margin of the Manawatu region is dominated by the Tararua–Ruahine Ranges, part of the North Island axial ranges which form the forearc region of the Hikurangi margin. The Tararua–Ruahine Ranges rise to over 1500 m above sea-level (Figs. 1 and 2), and are comprised of strongly deformed Mesozoic greywacke basement and marine cover beds uplifted during the Plio-Pleistocene (Stevens, 1990; Heerdegen and Shepherd, 1992).

West of the ranges lies the Plio-Pleistocene Wanganui Basin (Pillans, 1983; Naish and Kamp, 1995; Naish et al., 1998; Abbott and Carter, 1999). The Wanganui Basin formed in a subsiding back-arc position related to the modern Pacific–Australia plate boundary (Anderton, 1981). Prior to uplift through the late Quaternary,

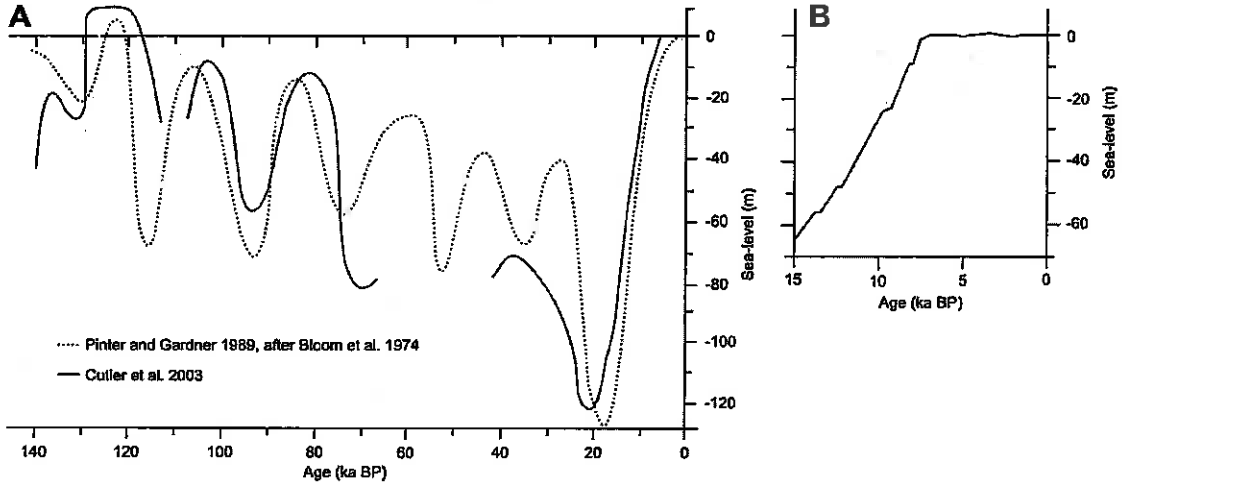

Fig. 3. Sea-level changes in the southwest Pacific region surrounding New Zealand as documented by records from (A) the Huon Peninsula, Papa New Ginuea, (B) the New Zealand region (after Carter et al., 2002).

sedimentation within the Wanganui Basin broadly matched subsidence rates. As a result, basin fill comprises numerous cycles of deep and shallow marine sedimentary successions related to glacio-eustatic sea-level cycles, representing one of the most complete and undeformed shelf and shallow marine Plio-Pleistocene records in the world (Pillans, 1983; Naish and Kamp, 1995; Naish et al., 1998; Abbott and Carter, 1999).

In the Manawatu region this Plio-Pleistocene record is represented by a series of low angle marine terraces formed by marine erosion during the last two interglacials (Heerdegen and Shepherd, 1992). Falling sea-levels leading into the LGM and tectonic uplift have resulted in the exposure of Last Interglacial marine deposits and the subsequent dissection of a seaward sloping raised marine terrace. Known locally as the Tokomaru Marine Terrace, this remnant of the Last Interglacial rises from 30 m above sea-level near Shannon, to over 90 m above sea-level near Palmerston North, reflecting a northerly increase in uplift rates along the western flanks of the Tararua–Ruahine Range. The terrace comprises soft sandstones, siltstones and gravels covered by 1–4 m of loess blown from exposed braided river beds by the prevailing northwesterly winds during the LGM. The Kawakawa Tephra (known locally as the Aokautere Ash), a volcanic ash associated with the Oruanui eruption dated at 26,600 ± 300 cal BP (Litchfield and Berryman, 2005) is present within the loess, providing a chronological control in the Manawatu area.

Continental shortening across the Wanganui Basin has resulted in the formation and development of a number of active, fault-controlled anticlinal structures which manifest at the surface and have influenced the geomorphic evolution of the coastal plain. Te Punga (1957) described five anticlines of 'prominent geomorphic expression' across the Manawatu Plain: Marton; Mt Stewart-Halcombe; Feilding; Oroua; and Pohangina (Fig. 2). Rich (1959) supplemented this list, adding the Himatangi anticline, together with an elongate depression, the Kairanga trough, which extends from Feilding township to Lake Horowhenua. A further anticline, the Shannon, was described by Hesp (1975) in the vicinity of Shannon township. Although Anderton (1981) considered that the faults underlying the anticlines were mostly active during the Pleistocene, Melhuish et al. (1996) showed that the growth of the Mt Stewart-Halcombe anticline has been ongoing since c. 3.1 Ma. However, only the westernmost fault underlying the Mt Stewart-Halcombe anticline has been active since c. 400 ka (Jackson et al., 1998). A prominent geomorphological characteristic of the four northernmost anticlines is the preservation of a mid-Quaternary (c. 300 ka) marine horizon across the folds, leading Jackson et al. (1998) to conclude that folding has been ongoing since at least this time.

3. Geomorphology of the Manawatu floodplain

The course of the Manawatu River across the Holocene coastal plain has been influenced by the Himatangi anticline and Poroutawhao 'High' (Fig. 2), which have anchored the mouth of the river at Foxton Beach and forced the course of the river below Rangiotu in a curve south, then east (Hesp, 1975; Stevens, 1990; Heerdegen and Shepherd, 1992). Directly downstream of the Manawatu Gorge, the Pohangina anticline directs the course of the Manawatu River to the southwest.

West of the Tararua–Ruahine Ranges the Manawatu River is characterised by two discrete channel phases (Figs. 1 and 2). Between the Manawatu Gorge and Opiki the river is gravel-bedded with a relatively steep slope (0.0012), and sinuosity of 1.4 (Page and Heerdegen, 1985). The floodplain here fits the classification of a medium energy non-cohesive wandering (B2) or lateral migration (B3) floodplain (Nanson and Croke, 1992), being comprised of gravels, sands and silts. This reach features a flight of four well defined aggradational river terraces evident between the Manawatu Gorge and Palmerston North City, all formed in response, at least in part, to changing base levels (Clement and Fuller, 2007). The highest river terrace (Forest Hill) in the lower Manawatu is associated with the (early?) MIS 4 (Otiran) cold stage (Heerdegen and Shepherd, 1992). Cutting of this terrace and refilling formed the Milson terrace, which Heerdegen and Shepherd (1992) suggest correlates with the Rata terrace in the Rangitkei (MIS 3?). The Last Glacial Maximum (LGM) floodplain is associated with the Ashhurst Terrace, correlating with the Ohakean terrace group in the Rangitikei (Clement and Fuller, 2007) and widespread valley floor aggradation in the eastern North Island (Litchfield and Berryman, 2005). The lowermost unit preserved (Raukawa) is of undetermined age, but presumed to be Holocene (Heerdegen and Shepherd, 1992).

From Opiki to the coast the river lacks the competency to transport gravel, being sand-bedded, with a sinuosity of 2.4 and a low gradient (0.0002; Page and Heerdegen, 1985). The modern floodplain in this lower reach reflects this lower energy, muddy-silty-sandy phase, being fine-grained in nature. However, in some bends lateral migration is evident, with well developed scroll patterns observed. This would classify the lower floodplain as a medium energy, lateral migration, scrolled/backswamp floodplain (B3b/c) (Nanson and Croke, 1992). River terraces are not evident in this part of the floodplain, although Hesp and Shepherd (1978) indicated that the floodplain is underlain at depth by gravels, with a radiocarbon age of c. 40,000 BP, possibly correlative with the Milson Terrace. This suggests the terraces plunge beneath the increasing thickness of fine-grained alluvium and estuarine deposits in the lowermost part of the valley as the river graded to a lower sea-level (Litchfield and Berryman, 2005).

4. Development of the Koputaroa dune field

Aeolian activity through the Last Glacial Maximum resulted in the formation of a parabolic dune field at Koputaroa (the Koputaroa Phase: Cowie, 1963), partially overlying the Tokomaru Marine Terrace (MIS 5e) and adjacent Ohakean-age aggradation surface (MIS 2). Koputaroa Phase dunes are regionally extensive, extending southwards to the vicinity of Otaki, there overlying an Ohakean-age aggradation surface. Cowie (1963) ascribed the Koputaroa Phase an age of 20–10 ka, based on stratigraphic relationship with the Oheakan surface, the presence of the Kawakawa Tephra (Cowie, 1964) interbedded within the dune sands, and palynological examination of peat from within the dunes which indicated that peat accumulation occurred under a much cooler climate than present. Both Cowie (1963) and Fleming (1972), who obtained a radiocarbon date of 35 ± 1.7 ka from peat at the base of a Koputaroa dune, considered it unlikely that the dunes were derived from a marine source, given that sea-level would have been much lower than present during the period of active dune formation. Instead, they suggested that the Koputaroa dune building phase was supplied locally from the beds of braided rivers.

Shepherd (1985) analysed the heavy mineral content and roundness of Koputaroa dune sand, modern fluvial sand and Holocene coastal dune sand and concluded that the Koputaroa dune sand was derived from a marine source. This conclusion was based primarily on the heavy mineral content of the Koputaroa dune sand, which comprises hornblende, hypersthene and augite. Shepherd (1985) considered that these minerals were originally sourced from the Taranaki and Central Volcanic regions to the north and subsequently transported to the coast by rivers before being carried southwards to the Manawatu by littoral drift.

In contrast, the sediment load of the rivers in the region of the Koputaroa Phase dunes consists of quartzo-feldspathic sediment derived from the Mesozoic greywacke which form the axial Tararua–Ruahine Ranges. Based on these sedimentological characteristics Shepherd (1985) therefore suggested that the most likely source of the Koputaroa Dunes was the mobilisation of transgressive and regressive deposits on the inner continental shelf that were deposited during relative higher sea-levels associated with interstadials of the last glacial cycle (MIS 5a, 5c and MIS 3) when sea-level rose to within 30 m of present levels (Fig. 3; e.g., Bloom et al., 1974; Chappell et al., 1996; Cann et al., 2000; Murray-Wallace, 2002; Cutler et al., 2003). During the interstadials marine sands would have been deposited only 15–20 km west of the location of the Koputaroa Dunes, providing sandy material that was subsequently reworked and transported landward during the cooler and windier conditions through the LGM.

However, while acknowledging the remoteness of the Koputaroa dune field to the LGM shoreline (approximately 30 km to the west), Shepherd (1985) ignores results which show an extreme similarity between Koputaroa phase dune sand and sand eroded from the Tokomaru Marine Terrace (MIS 5e). It is therefore important to make the distinction between the marine origin of the dune sands, as determined by Shepherd (1985), and the source of the dunes, supporting the assertions of Cowie (1963) and Fleming (1972) that the Koputaroa dunes are source-bordering dunes derived from large sand bars formed from sands eroded from the Tokomaru Marine Terrace that occupied the lower Manawatu River during this cold, dry, windy period (e.g., Nott and Price, 1991; Page et al., 1996; Bullard and McTainsh, 2003; Hesse et al., 2004).

More recently, Duller (1996) obtained nine infrared stimulated luminescence ages from Koputaroa dunes at Koputaroa, while Hawke and McConchie (2005) obtained three thermoluminescence ages from dunes in the vicinity of Otaki, all which range between 46 and 10 ka. The synthesis of geochronological data indicates that there were two phases of dune activity separated by a hiatus in aeolian transport most likely associated with the stabilisation of the Koputaroa Dunes (Shepherd, 1985; Shepherd and Price, 1990; Duller, 1996; Hawke and McConchie, 2005). Results from these studies show that the initial phase of deposition between occurred between 50 and 40 ka when sea-levels oscillated between 40 and 50 m below present levels (Fig. 3). At this time the dunes would have been only 5–10 km from the coast, which supplied sediment for this initial phase of development. At 50–40 ka, these ages are significantly older than the radiocarbon age obtained on the basal peat obtained by Fleming (1972). Duller (1996) postulates that this discrepancy may be the result of the peat sample being contaminated by modern carbon.

Following the initial phase of dune development active dune migration ceased until re-mobilisation occurred leading into and during the LGM, at which time the Kawakawa Tephra was incorporated into the dune successions. This phase of dune activity continued through the LGM until c. 11 ka (Muckersie and Shepherd, 1995; Duller, 1996; Hawke and McConchie, 2005). Dune activity during the period of 22–11 ka most likely reflects an increase in the intensity of westerly winds and a decrease in vegetation associated with drier and colder climate (Pillans et al., 1993). Through the LGM the water table would also have been substantially lower than present, resulting in less moisture at the surface to inhibit aeolian transport.

5. Holocene sedimentary successions: estuarine phase

The latter stages of the Holocene marine transgression inundated the lower Manawatu valley, resulting in an estuary that extended east to Shannon and north to Opiki (Hesp and Shepherd, 1978; Heerdegen and Shepherd, 1992). At this time the Himatangi and Marton anticlines had attained sufficient altitude, together with the Poroutawhao 'High', a bedrock dome extending 5–24 m above mean sea-level, preventing major marine incursions into the incised-valley (Te Punga, 1953; Rich, 1959; Hesp, 1975; Hesp and Shepherd, 1978; Heerdegen and Shepherd, 1992).

The timing of estuary formation in the lower Manawatu has previously been presumed to coincide with the stabilisation of sea-levels c. 6700 cal yr BP, following Gibb's (1986) Holocene sea-level curve for New Zealand. However, recent research from the eastern seaboard of Australia indicates a much earlier culmination of the Holocene marine transgression c. 7500 cal BP (e.g., Horton et al., 2007; Sloss et al., 2007; Woodroffe, in press). Most recently, Clement et al. (2008a,b), in a New Zealand-wide review of indicators of Holocene palaeo sea-level (Fig. 4), concluded that Gibb (1986) most likely underestimated the age of attainment of present mean sea-level (PMSL) following the marine transgression, showing that New Zealand sea-levels rose from −5 m PMSL c. 8500 cal BP to +0.3 m above present mean sea-level (PMSL) by 7550 cal BP (Fig. 5). However, while this sea-level history demonstrates much similarity with curves from eastern Australia, it suffers, as does Gibb's (1986) work, by virtue of being a regional sea-level curve reconstructed from a relatively distant source area and using a variety of proxy sea-level indicators with variable relationships to their contemporary sea-level (Fig. 6).

Shells recovered by Shepherd (1987) place the upper limit of the estuary in the vicinity of Murray's Farm (Fig. 1). Here, shells from the top of the estuarine facies suggest that estuarine infill was succeeded by fluvial progradation c. 6780 ± 480 cal BP. The eastern margin of the palaeo estuary features a number of box-shaped valleys in the vicinity of Shannon township, formed when small river valleys incised into the Tokomaru Marine Terrace during the LGM were filled with estuarine sediment. Here, shells 1.1 m above PMSL place the date of succession from estuarine to fluvial deposition at 6810 ± 175 cal BP (Hesp and Shepherd, 1978). This indicates that sea-level was close to present levels well before 7000 cal BP, to allow for the development of a low-energy estuarine environment within the incised-valley close to present sea-level by 6800 cal BP.

6. Holocene sedimentary successions: coastal dunes

During the late Holocene, following the culmination of the Holocene marine transgression, sand which had migrated landward with rising sea-levels was transported to the nearshore, beach and back-beach environment by the dominant westerly swell, and further inland by prevailing west–northwesterly winds to form the most extensive transgressive dune field in New Zealand. The dune field covers c. 900 km^2 and consists largely of stabilised parabolic dunes aligned almost exactly parallel with the dominant onshore WNW winds. The dunes have migrated inland between the Wanganui and Manawatu Rivers over the inner margins of the relatively flat Holocene coastal plain to reach the floodplains of the Manawatu. The dune field also extends south into the Otaki-Te Horo area, but here the dunes are much more subdued and restricted in their landward progradation. The prograding dunes have frequently blocked the valley of the small streams which flow westwards across Pleistocene marine terraces, forming small lakes and swamps parallel with the coast (Fig 1).

The Manawatu coast is particularly suited for aeolian transport of fine sand and the mobilisation of large parabolic and transgressive dunes due to: (1) the favourable onshore wind and wave regime. The Manawatu coast is dominated by onshore west–northwest winds strong enough to initiate sand transport approximately 33% of the time (Fig. 1; Shepherd, 1987; Muckersie and Shepherd, 1995).

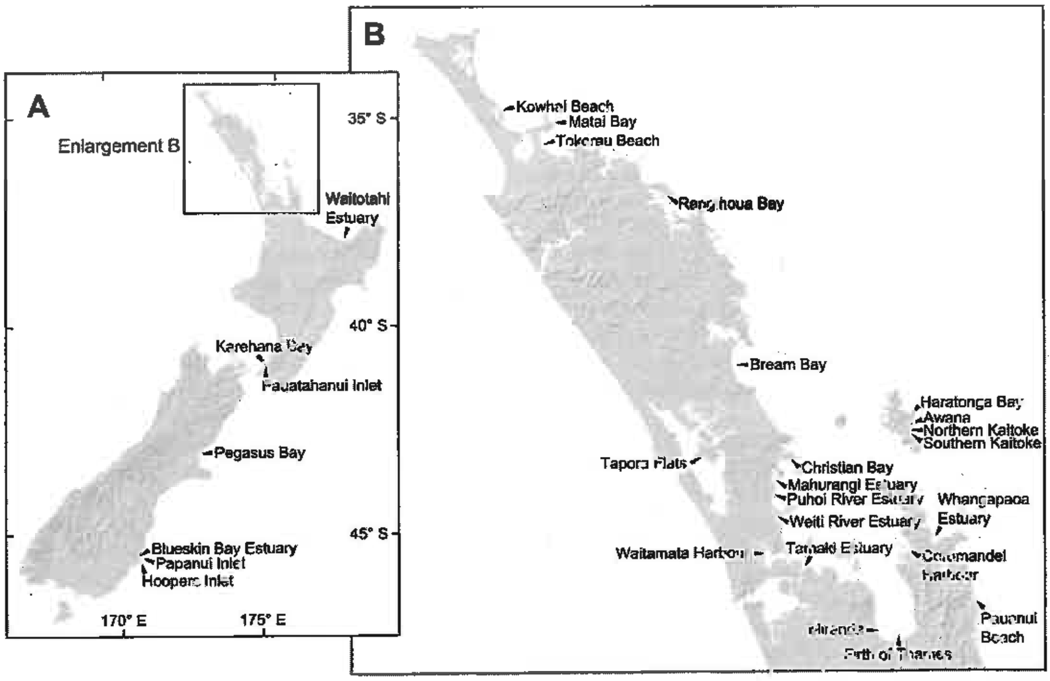

Fig. 4. Map of the spatial distribution of sea-level index points used to compile the revised Holocene sea-level curve for the New Zealand region (Fig. 5).

(2) The dominant onshore wave-approach maintained on the Manawatu coast by the dominant west–northwest wind direction. (3) A large supply of sandy sediment from the Wanganui and Whangaehu Rivers in the north of the Wanganui Basin and the Rangitikei and Manawatu Rivers further south (Gibb, 1977). Together, these four rivers supply over 10 million tonnes of sediment to the Manawatu coast every year (Hicks and Shankar, 2003). (4) Manawatu beaches typically feature modally dissipative profiles dominated by fine-grained sand with wide gently sloping foreshores resulting in a large surface area with the potential for sediment mobilisation and aeolian transport.

Pedological studies by Cowie (1963) identified three dune building phases. More recent research based on radiocarbon and luminescence dating from deposits associated with dunes supports the concept of dune phases (e.g., Shepherd, 1987; Muckersie and Shepherd, 1995; Hawke and McConchie, 2006). However, these later results indicate that the phases were less distinct than originally proposed. McFadgen (1985) correlated the latter two phases with depositional episodes from throughout the North and South Islands, therein subdividing the Waitarere dunes into Early and Late Phase.

6.1. Foxton Phase (6.5–1.6 ka)

The Foxton Phase of Holocene dune migration is the most extensive with dunes extending up to c. 16 km inland and rise to a maximum elevation of c. 30 m above the surrounding alluvial plain. Cowie (1963) estimated, based on the degree of soil

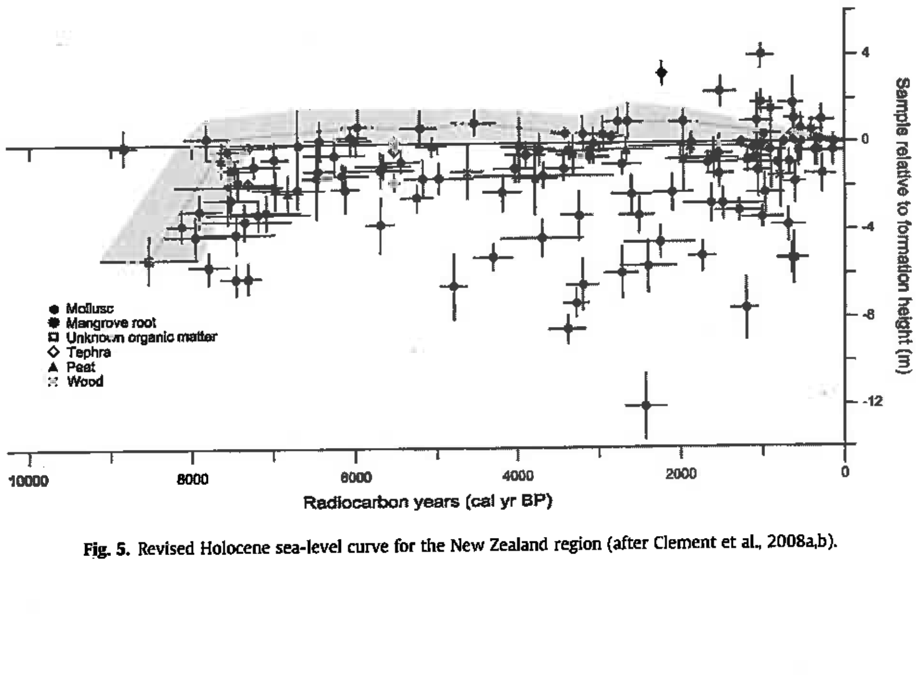

Fig. 5. Revised Holocene sea-level curve for the New Zealand region (after Clement et al., 2008a,b).

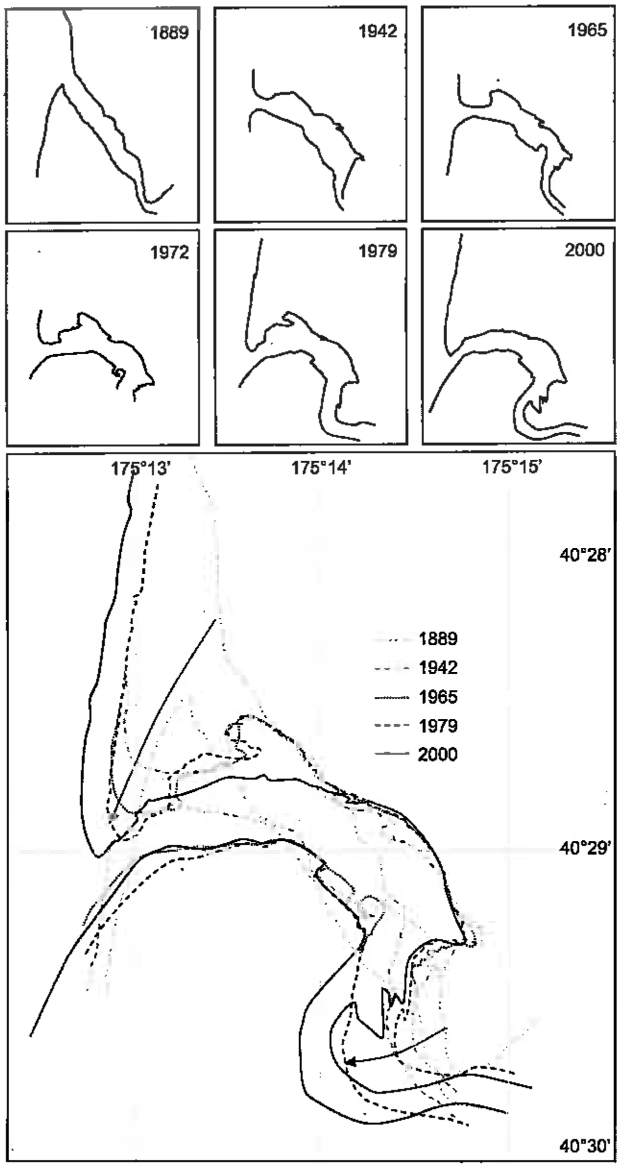

Fig. 6. Time-series map of recent geomorphological change of the modern Manawatu River estuary. 1972 is omitted from the overlay to enhance legibility given the relatively little change in profile from 1965 to 1972.

development, that the Foxton Phase of dune development was initiated c. 4 ka, continuing until c. 2 ka. The dunes are characterised by having a moderately well developed soil profile with a c. 30 cm thick organic-rich "A" horizon and a distinct dark brown "B" horizon of variable depth.

More recent research suggests that the active dune migration was initiated soon after the culmination of the Holocene marine transgression, continuing until c. 1.6 ka (Shepherd, 1987; Shepherd and Price, 1990; Muckersie and Shepherd, 1995; Hawke and McConchie, 2005). During the early stages of the Holocene marine transgression sediment stored on the continental shelf would have been moved shoreward by rising sea-levels providing a sediment source for the formation foredunes in the back-beach environment (Gibb, 1986; Shepherd, 1987). At this time the Manawatu coastline was approximately 4 km landwards of its current position (Muckersie and Shepherd, 1995; Hawke and McConchie, 2006).

By 2.3 ka the dunes had prograded over the Himitangi anticline, reaching the floodplain near the junction of the Manawatu and Oroua Rivers and leaving in their wake a thin aeolian veneer over the floodplain. The Foxton Phase dunes reached their maximum extent c. 16 km from the present day coastline, stabilising c. 1.6 ka (Shepherd, 1987). Muckersie and Shepherd (1995) have shown that there was a possible hiatus of active dune formation at the coast that lasted until c. 3.5 ka while the already mobilised dunes continued to migrate inland. Following this hiatus the Foxton Phase resumed, during which time the Taupo Pumice (1717 cal BP: Lowe et al., 2008) was incorporated into the active dunes (Muckersie and Shepherd, 1995).

Based on sedimentological analysis and eight OSL ages obtained on the Foxton Phase dunes at Otaki-Te Horo, Hawke and McConchie (2006) named the earlier advancement of the Foxton Phase as the Swamp Road dunes. This distinction is based on differing sedimentary characteristics: Swamp Road dunes feature more poorly sorted sediments with a higher magnetic content than Foxton Phase dunes, due to the higher energy conditions of the gravel beaches and gravel-bedded Otaki River.

The OSL ages obtained from Swamp Road led Hawke and McConchie (2006) to conclude that while the Swamp Road dunes are of similar age to the Foxton Phase dunes (c. 4000 years old in the Otaki-Te Horo area), Swamp Road represents the leading edge of the Foxton Phase, and is locally influenced by sediment supply from the Otaki River and the gravel beach. Additional OSL ages from the Foxton dunes in this area rage from 1830 ± 540 years to 2980 ± 1070 years, consistent with the age of landward Foxton Dunes in the region of the Manawatu River (Muckersie and Shepherd, 1995).

6.2. *Motuiti Phase (1–0.5 ka)*

The dunes and sand plains of the Motuiti Phase extend up to 11 km from the present day coast. North of the Manawatu River a line of small lakes mark the contact between the Motuiti dunes and the younger Waitarere Phase (seaward) dunes (Fig. 1; Cowie, 1963). Dunes of the Motuiti Phase are characterised by a shallow (c. 15 cm) dark grey / black organic stained sandy "A" horizon and a c. 18 cm pale yellowish brown "B" horizon which overlays grey sand. The Motuiti Phase of active dune migration is believed to have been initiated c. 1 ka, based on the radiocarbon age determination of 850 ± 50 years of an *in situ* tutu (*Coriaria* spp.) stump rooted in the underlying (preceding) Foxton Phase and overlain by 22 m of the Motuiti Phase. Based on its stratigraphic position Cowie (1963) suggested that the tutu was killed by advancing sand. A similar chronology for the Motuiti Phase dunes was also reported from the Otaki area, where an OSL age of 670 ± 140 years was obtained on dune sands (Hawke and McConchie, 2006). Based on the timing of active dune formation this phase of active dune mobilisation has been attributed to de-vegetation of previously stabilised dunes associated with Maori occupation which began in the Manawatu region 650–700 years ago (McGlone, 1983; McGlone and Wilmhurst, 1999).

6.3. *Waitarere Phase (<120 years)*

The Waitarere Phase dunes form a coastal belt of active and stabilised dunes that extend inland between 0.5 and 4 km (Fig. 1; Cowie, 1963). These dunes lack soil development apart from a slight darkening of the surface few centimetres by organic matter and sands and are relatively unweathered. The Waitarere Phase of active dune mobilisation is <120 years old, as Waitarere Phase dunes have prograded over European artefacts and introduced plants (Cowie, 1963). Accordingly, this phase of dune activity has been partly attributed to overgrazing and the burning of original vegetation on previously stabilised dunes (Cowie, 1963). Vegetated dunes of the Waitarere Phase are extensively covered with *Scirpus nodosus* (Pingao), *Spinifex hirsutus* (Spinifex) and *Ammophila arenarua* (Marram). However, there are large areas that are occupied by active foredune blowouts and parabolic dunes. Since active dune progradation continues today despite significant human intervention in the form of planting, dune re-shaping and fertilising, it may be that early human activity merely aided natural processes in the development of the most recent phase of dune activity. Supporting evidence that active dune migration was dominantly a natural process is provided by an OSL age of 470 ± 130 years on dunes from the Otaki-Te Horo area (Hawke and McConchie, 2006) which pre-dates European occupation.

7. Contemporary coastal progradation

The Manawatu River has undergone significant geomorphological changes over the last 120 years. Based on historic maps from the region it has been possible to map these geomorphological changes (Fig. 6). The most significant change occurs at the mouth of the river estuary which in 1889 was much further north than its present location. A combination of westerly swell waves off the coast and onshore WNW winds, high sediment loads and a dominantly southward long-shore drift has resulted in a progressive southward movement to the mouth of the river estuary that is associated with the formation of a sand spit. However, it is believed that this gradual movement southward is not permanent. This is because the spit has formed over the path of previous river channels and there is a high probability that during extreme flood events the river will attempt to return to a more northerly path.

At the mouth of the Manawatu River the sand dunes that occur on the spit are accreting inland (Fig. 1). This can be seen by the development of blow-out features in the foredune and the development of parabolic dunes. Both the spit and the dunes have formed within the last 120 years, which is testament to how active the area is. The growth of the spit over the last 120 years has resulted in the formation of mudflats and salt marshes in the back-barrier environment (Fig. 6).

8. Comparisons of lower Manawatu valley stratigraphy to tripartite facies models of incised-valley infill

The sediment fill of drowned incised-valleys is complex, reflecting the interplay between fluvial and marine processes in the marginal marine environment The application of facies analysis and sequence stratigraphic principles to estuarine deposits has contributed to a number of widely recognized "tripartite" facies models, which broadly differentiate between wave- and tide-dominated estuarine systems (e.g., Roy, 1984a,b; Dalrymple et al., 1992; Allen and Posamentier, 1993, 1994). These models laid the foundation for detailed investigations of the characteristics of the sedimentary infill of incised-valleys associated with transgressive and highstand conditions. Such infill processes manifest distinct bio-lithological facies associations suitable for the reconstruction of palaeo-depositional environments and sea-level histories (e.g., Heap and Nichol, 1997; Sloss et al., 2005, 2006a,b, 2007, 2010; Kennedy et al., 2008; Vis et al., 2008).

These established tripartite facies models recognize three to four broadly similar sedimentary environments. All models identify a fluvial/floodplain depositional environment including deltaic systems prograding into central receiving basins. For example, Dalrymple et al. (1992) recognize a fluvial delta environment at the head of the estuary, while Allen and Posamentier (1993, 1994) include a shoreface environment. The central portion of a drowned incised-valley system is occupied by a central basin facies consisting of fine-grained organic-rich silty mud deposited in a low-energy back-barrier environment. At the seaward end of the system the tripartite facies models identify marine-influences sediments, including barrier spit, back-barrier sand-flats, washover deposits, flood-tide delta and tidal inlet channel deposits.

The emphasis tripartite facies models place on the interplay between rising sea-levels, sediment supply and the antecedent topography of the lowstand incised-valley is critical. Where the earlier work by Roy, Dalrymple and others has a strong emphasis on geomorphology, more recent work has a much stronger stratigraphic component (e.g. Heap and Nichol, 1997; Sloss et al., 2006a,b, 2007, 2010; Wilson et al., 2007). For example, the stratigraphy of an incised-valley will record the interaction between the rate at which accommodation space is generated and the sedimentation rate. Therefore, transgressive facies will dominate valley fill where the relative sea-level rise exceeds the rate of sediment flux. Regressive facies prevail where the converse occurs (e.g. Curray, 1964; Boyd et al., 1992; Heap and Nichol, 1997; Sloss et al., 2006a,b, 2007, 2010).

Under conditions of rising sea-level the tripartite models of valley fill predict that fluvial, estuarine basin and barrier environments will migrate landward, at which point, once the sea-level highstand is attained, the estuary progresses from an immature system with a larger amount of accommodation space to a more mature system that is largely infilled by fluvial sediment (Roy et al., 1980; Roy, 1994; Heap et al., 2004).

Following the model presented by Roy et al. (1980) the lower Manawatu River valley is anogalous to a drowned river valley estuary: a deeply incised-valley with a wide deep-water entrance. Under the Dalrymple et al. (1992) model, the Manawatu valley would classify as a wave-dominated estuary. As Wilson et al. (2007) note, the case studies of Holocene incised-valley infill upon which the tripartite facies models are based commonly come from estuaries of extensive size (40–100 km in length) on stable coastlines, with comparatively limited sediment supplies (e.g., Allen and Posamentier, 1993, 1994) and low river discharges (e.g., Dalrymple et al., 1992). However, as the Manawatu valley is sited in a tectonically active landscape and is subject to large sediment inputs and river flows, it is therefore prudent to consider the extent to which the estuary evolution in Manawatu follows the tripartite evolutionary models.

While the major source of sediment for mid-Holocene dune development adjacent to the palaeo lower Manawatu incised-valley is commonly cited as transgressive sands moved onshore by rising sea-levels during the early Holocene, neither Hesp (1975) or Hesp and Shepherd (1978) record the presence of a sedimentary unit anogalous to these transgressive sands in the lower Manawatu valley. On the southeast coast of Australia Sloss et al. (2005, 2006a,b, 2007) have reported extensive transgressive sand sheet units overlying the antecedent late-Pleistocene land surface in shallow incised-valleys. These transgressive sand sheets were deposited inside the valleys when rising post-glacial sea-levels breached Last Interglacial barriers at their mouths. The absence of similar sizeable transgressive units within the lower Manawatu valley may be due to the presence of the Porotawhao High and Hiitangi Anticline at the mouth of the valley, restricting marine inundation into the valley during the early stages of sedimentary infill, resulting in any transgressive facies being spatially restricted to the mouth of the valley (e.g., Roy et al., 1980, 2001; Chapman et al., 1982; Roy, 1984a,b, 1994; Dalrymple et al., 1992). Another factor that would have resulted in restricting the transgressive phase of sedimentation is the high flow regime and sediment load of the Manawatu River (maximum flows of 4500 cumecs; current sediment yield is 3.4 Mt yr^{-1}) which would have removed, reworked or otherwise obscured the transgressive facies by fluvial processes, with the infill system essentially operating as a depositional environment experiencing a forced regression.

Increasingly, studies of incised-valley fill in New Zealand demonstrate that the development and preservation of the central basin facies is affected by the combined effects of the allocentric controls of accommodation space, tectonism, sedimentation rate and sea-level behaviour (e.g., Heap and Nichol, 1997; Wilson et al., 2007; Abrahim et al., 2008; Kennedy et al., 2008). In Tamaki Estuary, Auckland, Abrahim et al. (2008) report the absence of a central basin facies in a sequence that is otherwise broadly consistent with the idealised case of a wave-dominated incised-valley system. In a study from the narrow Pakarae River Estuary (<500 m wide), on the tectonically active Raukamarua Peninsula, East Coast North Island, Wilson et al. (2007) document poor preservation of central basin sediments. Here, this is attributed to rapid fluvial progradation into Pakaerae Estuary in direct response to coseismic uplift events, leading to the truncation of central basin deposits in the sedimentary sequence. Both Heap and Nichol (1997) and Kennedy et al. (2008), working in Weiti River Estuary and Whanganui Inlet respectively, report stratigraphic records where high sedimentation rates, shallow valley form and lack of a seaward barrier have limited the development of a low-energy central basin.

In the Manawatu the deposition of estuarine sediment into the central basin was extremely short lived. Though the lower Manawatu valley is not particularly narrow, or prone to coseismic uplift events, it does posses an extremely high sediment load, echoing the situation described by Heap and Nichol (1997) and Kennedy et al. (2008). Notably, both Hesp (1975) and Hesp and Shepherd (1978) document an abrupt transition from estuarine tidal flats comprising blue–grey silt and clay (equivalent to the central basin facies) to alluvium; neither describe an intermediary bay-head delta sequence. The absence of bay-head delta facies may indicate reworking by fluvial processes. Alternatively, rapid and high rates of sediment delivery by the Manawatu River may have denied bay-head delta formation, as swift infill of the central basin eliminated the necessary accommodation space, thereafter leading to sediment bypass.

9. Conclusions

The present geomorphology of the Manawatu River, associated floodplain, coastal plain and coastal landscape has been controlled by sea-level fluctuations over the last full glacial cycle and the influence of such changes on the antecedent Plio-Pleistocene landscape. During the Last Glacial Maximum, when sea-level was c. 120 m lower than it is today, the Manawatu River incised a broad, deep valley, while many small tributaries dissected MIS 5e marine terraces flanking the inland margins of the region. Cold, windy conditions also initiated The Koputaroa phase of dune development, sediment being sourced from the large, exposed bars of a then-braided Manawatu River.

During the Holocene post-glacial marine transgression rising sea-levels inundated the incised-valley c. 7500 cal BP. During the latter stages of the post-glacial marine transgression natural barriers associated with the Himatangi anticline and Poroutawhao High restricted oceanic influences, possibly preventing transgressive sand from encroaching into the incised-valley, while facilitating the formation of a low-energy estuarine environment and the partial infilling of the incised-valley with fine-grained estuarine mud. The partial infilling with estuarine sediments within the dissected terraces resulted in flat valley floors and box-shaped valleys. Following the culmination of the post-glacial marine transgression and during the Holocene sea-level highstand rapid sedimentary infill occurred within the incised-valley associated with high volumes of fluvial sediment introduced by the Manawatu River. This resulted in estuarine deposition being quickly succeeded by fluvial deposition and floodplain aggradation, creating an extensive coastal plain. Rapid infill and consequent sediment bypass may account for the lack of characteristic bay-head delta development. These findings support results presented for other New Zealand incised-valley systems subject to large sediment loads. Thus, it is becoming increasing clear that tripartite models of incised-valley evolution developed for stable coasts are not representative of New Zealand estuarine systems.

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