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The Late Quaternary geomorphology of the Manawatu coastal plain

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Regional Setting

the Manawatu-Wanganui region is located in the southwest of the north island of New Zealand. The region stretches from Himatangi in the south to just south of Mangaweka in the north, and from the west coast to the summit of the Tararua Ranges to the east. The Manawatu region currently has a temperate maritime climate. Rainfall increases from 800 mm at the coast to more than 5000 mm along the ranges with no marked seasonality (Heerdegen and Shepherd, 1992). Sediment is deposited at the coast by local rivers, the largest of which are the Wanganui, Rangitikei and Manawatu Rivers (Fig.1). Waves approaching the coast of the Manawatu region come from either the southwest or the west. Waves approaching from the southwest are generated in the northern Cook Strait and have a maximum fetch of 100 km resulting in moderate wave energy. Waves approaching from the west (Tasman Sea) have a greater fetch resulting in a dominant westerly swell. North of the Wanganui River the westerly swell waves approach the coast at an oblique angle resulting in a southward moving longshore drift. Accordingly, waves in the nearshore not only transport large quantities of sand-sized sediment shoreward from the northwest but by the southwards as well.

North of the Wanganui River the sediment is dominated predominantly by volcanic detritus (hornblende, hypersthene and augite) originally sourced in the Taranaki and Central Volcanic regions to the north and subsequently transported to the coast by rivers and carried southwards by dominant littoral drift (Gibb, 1977). The present coastline between the Wanganui River and the Otaki River is dominated by fine-grained sandy sediments derived from the greywacke ranges to the east. The fine-grained nature of the sediment combined with moderate wave energy with shorter periods (< 10 s) and a relatively low tidal range has resulted in a coast characterised by lower gradients and multiple sand bars. Studies at Wanganui (Shand *et al.*, 2001), found that these bars form on the lower beach, then migrate seaward across the surf zone, and finally disappear several years later. The beaches of the Manawatu region tend to have lower cross-shore slopes and have modally intermediate (ripped) to dissipative (uniform) states (configurations). The beaches are backed by a foredune which may reach several metres in height. The form of the foredune varies alongshore from well vegetated stabilised incipit and established foredunes to poorly vegetated blown-out and hummocky dunes. However, south of the Otaki River the beaches are composed of a mix of fine-grained sand and gravels of fluvial origin. The result is a narrower, steeper and reflective beach profile with distinct cusps and berms and often with a less developed foredunes.

Geological setting

The eastern margin of the Manawatu is dominated by the Tararua-Ruahine Range which rise to over 1500 m. The ranges comprise strongly deformed Mesozoic greywacke basement and marine cover beds that were uplifted during the Plio-Pleistocene (Stevens, 1990; Heerdegen and Shepherd, 1992). West of the range lies the Plio-Pleistocene Wanganui Basin (Fig. 1; Pillans, 1983; Abbott and Carter, 1994; Naish and Kamp, 1997; Naish *et al*., 1998). This is a *c*. 40,000 km^2 sedimentary basin extending from Mt Taranaki in the north to the axial ranges of the North Island near Kapiti Island to the south. The basin formed in a subsiding back-arc position related to the modern Pacific–Australia plate boundary (Anderton, 1981). Sedimentation within the Wanganui Basin broadly matched subsidence throughout much of its history until the basin underwent uplift in the late Quaternary. As a result, basin fill comprises numerous cycles of deep and shallow marine sedimentary successions related to glacio-eustatic sea-level cycles and represents one of the most complete and undeformed shelf and shallow marine Plio–Pleistocene records in the world (Pillans, 1983; Abbott and Carter, 1994; Naish and Kamp, 1997; Naish *et al*., 1998). In the Manawatu region this Plio-Pleistocene record is represented by a series of low angled benches created by marine erosion and retaining a veneer of marine sediment that flank the western margin of the Tararua-Ruahine Range. These marine terraces represent marine inundation during interglacials associated with the lower Waitotaran, Upper Waitotaran/lower Nukumaruan, Upper Nukumaruan and the Lower Castlecliffian (Heerdegen and Shepherd, 1992).

Falling sea levels leading into the Last Glacial Maximum (LGM) and tectonic uplift resulted in the exposure of Last Interglacial marine deposits and the subsequent dissection of a seaward sloping raised marine terrace. Remnants of the Last Interglacial (Oturi; *c*. 128-115 ka) are well preserved in the Manawatu district forming broad planar surfaces flanking the adjacent Tararua-Ruahine Range known as the Tokomaru Marine Terrace (Fig. 1; Heerdegen and Shepherd, 1992). The terrace comprises soft sandstones, siltstones and gravels. Between 1 and 4 m of loess blown in from river beds by the prevailing northwesterly winds during the LGM cover the terraces. The Kawakawa Tephra, a volcanic ash associated with the Oruanui eruption dated at 22,590±230 yr BP (Wilson *et al*., 1988; locally known as the Aokautere Ash) is present within the loess, providing a chronological control in the Manawatu area. To the south of the Manawatu Gorge the narrow Tokomaru Marine Terrace rises to an elevation of 30 m near Levin to *c*. 90 m near Palmerston North, reflecting a northerly increase in uplift rates along the western flanks of the range (Heerdegen and Shepherd, 1992).

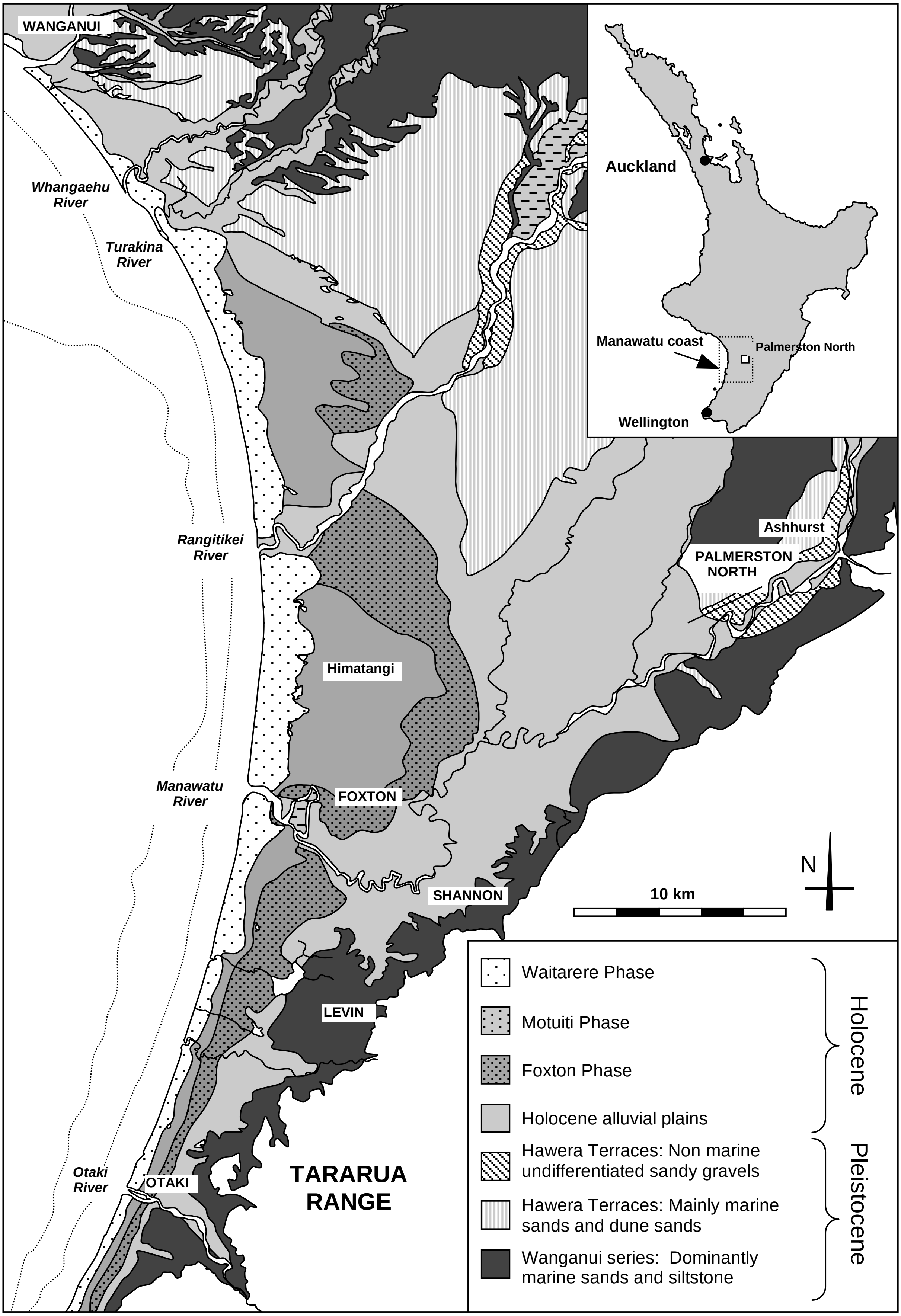

Figure 1: Quaternary geology of the Manawatu region (After Shepherd, 1987; Geological Map of New Zealand, 1:250 000, Sheet 10 (Wanganui), Sheet 11 (Dannevirke) and Sheet 12 (Wellington).

Manawatu floodplain

The lower Manawatu River is characterised by two discrete channel phases, and the floodplain characteristics reflect this. Between the Manawatu gorge and Opiki the channel has a relatively steep slope (0.0012), is gravel-bedded and has a sinuosity of 1.4 (Page and Heerdegen 1985). The floodplain associated with the Manawatu River comprises gravels, sands and silts and fits the classification of a medium energy non-cohesive wandering or lateral migration floodplain (Nanson and Croke 1992). Alluvium in this reach of the Manawatu River floodplain has four well defined terraces evident between Palmerston North and the gorge. These terraces have all been formed during the last glacial cycle, inset against the Tokomaru Marine Terrace, which represents the coastline during the last interglacial (OIS 5e). The highest river terrace (Forest Hill) in the lower Manawatu is associated with the (early?) 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 (OIS 3?). The Last Glacial Maximum floodplain (OIS 2) is associated with the Ashhurst Terrace, and the lowermost unit (Raukawa) is of undetermined age, but is presumed to be Holocene in age (Heerdegen and Shepherd 1992). The Holocene and modern alluvium retain a series of palaeo-meanders indicating a recent change in sinuosity of this reach (Page and Heerdegen, 1985).

From Opiki to the coast the river is no longer competent to transport gravel and is sand-bedded, with a sinuosity of 2.4 and a low gradient (0.0002; Page and Heerdegen 1985). In this lower reach the modern floodplain comprises muddy and silty sand, reflecting a transition from a high-energy upper reach to the more moderate-energy phase observed in the lower Manawatu River. In some meanders lateral migration is evident, with well developed scroll patterns observed. Thus, the lower Manawatu River can be classified as classified as a medium energy, lateral migrating, scrolled/backswamp floodplain (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 ^{14}C age of ~40 ka 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. Valley gradients were steeper at times of lower sea level.

The Koputaroa Dune Phase

An episode of aeolian activity during the LGM resulted in the formation of parabolic sand dunes near Koputaroa and Levin in the Horowhenua County (Fig. 1; Cowie, 1963; Fleming, 1972; Shepherd, 1985; Duller, 1996; Hawke and McConchie, 2005). The Koputaroa dunes lie on the Ohakean aggradations surface (OIS2) in a relatively restricted area just to the north of Levin. It was suggested that a marine source for the dunes was unlikely as sea level would have been much lower than present during active dune formation. Both Cowie (1963) and Fleming (1972) suggested that the Koputaroa Phase was related to the incisions of braided rivers below aggraded marine terraces resulting in sediment being available for transport into source bordering dunes. Fleming (1972) reported a radiocarbon age of 35,000±1700 yr BP obtained on peat towards the base of a Koputaroa Phase dune, indicating that initial aeolian deposition occurred during OIS3. Fleming also reported that at least two Koputaroa Phase dunes contained an interbedded horizon of Kawakawa Tephra indicating that aeolian activity continued into the LGM (OIS2). Investigation into the pollen within the interbedded peat layers within the dunes indicated that the dunes formed in a cool climate, also suggested active dune deposition during the LGM (Fleming, 1972).

Research by Shepherd (1985) and Shepherd and Price (1985) indicated that the Koputaroa Phase dunes have a coastal origin. Evidence supporting a marine source is based on the heavy mineral content comprising hornblende, hypersthene and augite. These minerals were originally sourced from the Taranaki and Central Volcanic regions to the north and subsequently transported to the coast by rivers and carried southwards into the Horowhenua district by dominant littoral drift. 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 range (Shepherd, 1985). Based on the sedimentological characteristics Shepherd (1985) suggested that the most likely source of the Koputaroa Dunes would have been the mobilisation of transgressive and regressive sediments deposited on the inner continental shelf during relative higher sea levels associated with interstadials of the last glacial cycle (OIS5a, 5c and OIS3 *c*. 37,000 years ago) when sea level rose to within 30 m of present levels (Chappell *et al*., 1996; Cutler *et al*., 2003; Bloom *et al*., 1974; Cann *et al*., 2000; Murray-Wallace, 2002). During the interstitials marine sands would have been deposited only 15 – 20 km west of the location of the Koputaroa Dunes and provided sandy material that was reworked and transported landward during the cooler and windier conditions leading into and during the LGM.

Recent research has confirmed the marine source of these dunes and provided additional chronological control on the timing of dune activity. Shepherd and Price (1985) also obtained a thermoluminescence (TL) age from the dunes immediately beneath the Kawakawa Tephra (*c*. 22,000 yr BP) and above the basal peat layer (dated at 35,000±1,700 yr BP; Fleming, 1972) that yielded an age of 24,200±3,700 indicating that the Koputaroa dune building phase was active leading into the LGM.

Duller (1996) obtained nine infrared stimulated luminescence ages and Hawke and McConchie (2005) obtained three TL ages from the Koputaroa Dunes that raged between 46,000 yr and 10,000 years. 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, 1985; Duller, 1996; Hawke and McConchie, 2005). Their results show that the initial phase of deposition occurred between 50,000 and 40,000 yrs (OIS3) when sea levels oscillated between 25 and 50 m below present levels and the dunes would have been only 5 – 10 km from the coast. The proximity to the coast during the interstitials would have provided the sediment supply for the initial deposition. However, 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 and that 1% contamination on an infantry old sample would give an approximate age of 35,000 yr B.P.

Following the initial deposition there was no active dune migration until the second phase of dune activity when the dunes were re-mobilised leading into and during the LGM and at which time the Kawakawa Tephra was incorporated into the dune successions. This phase of dune activity continued through the LGM until c. 10,000 years ago (Shepherd and Muckersie, 1995; Duller, 1996; Hawke and McConchie, 2005). Dune activity during the period of 22 – 10 ka most likely reflects a increase in the intensity of westerly winds and a decrease in vegetation associated with a drier and colder climate during the LGM and continued until becoming stabilised c. 10 ka (McIntyer, 1963; Pillans, et al., 1993; Thom et al., 1994; Duller, 1996; Hawke and McConchie, 2005). Leading into and during the LGM the water table would have also been substantially lower than present resulting in less moisture at the surface to inhibit aeolian transport.

Holocene sedimentary successions: Estuarine successions

During the LGM fluvial activity resulted in valley incision into the marine terraces, particularly those between Palmerston North and Levin, resulting in small river valleys and a moderately dissected landscape. During the Holocene marine transgression rising sea levels inundated the incised valleys in the lower Manawatu and Oroua Valleys (Fig. 2; *c*. 6,500 years ago; Gibb, 1987).

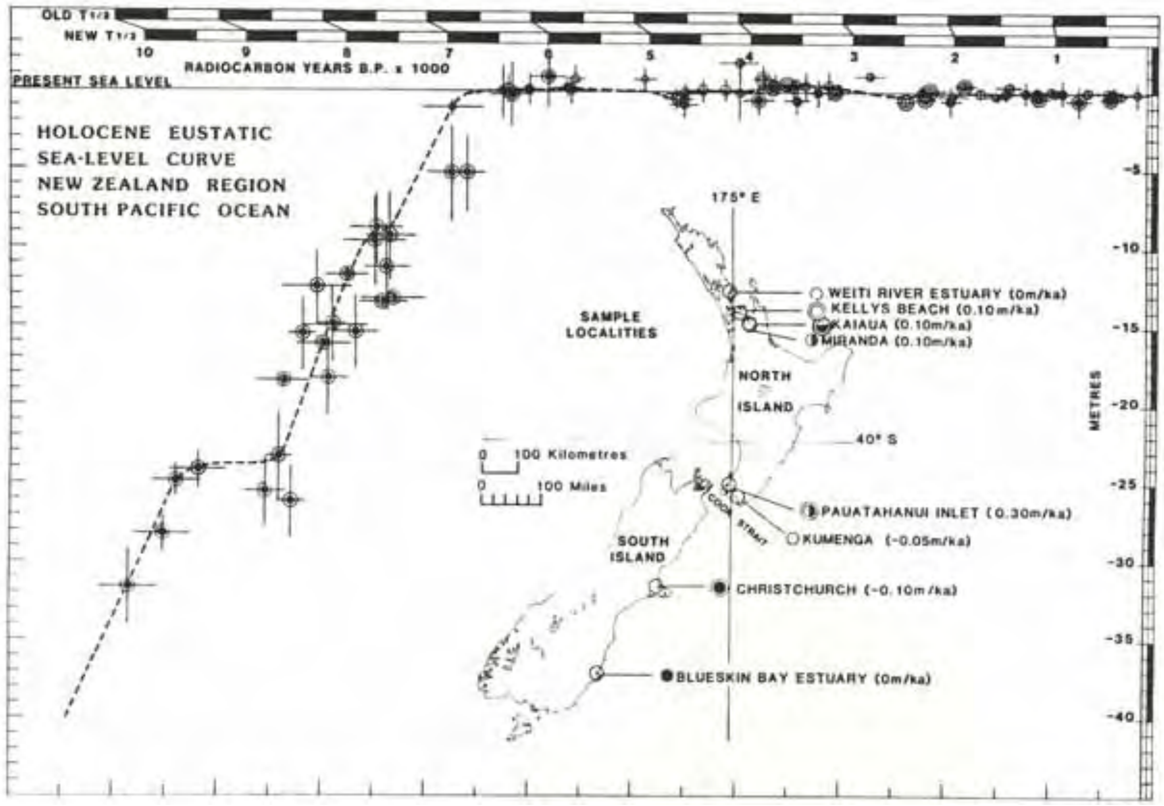

Figure 2: A New Zealand sea-level curve following the last glacial maximum and during the Holocene sea-level highstand established for New Zealand (from Gibb, 1987).

During the latter stages of the Holocene marine transgression natural barriers associated with the Marton and Himatangi anticlines and Poroutawhao High (Te Punga, 1975) restricted oceanic influences and facilitated the formation of low-energy estuarine environments and the partial infill of the incised valleys with fine-grained estuarine mud and alluvial plain aggradation (Fig.3, Hesp and Shepherd, 1978; Shepherd, 1987). Radiocarbon dating of estuarine shells preserved in the transgressive deposits place the time of estuarine formation close to the culmination of the PMT (6,330±70 yr BP; Hesp and Shepherd, 1978; Shepherd, 1987). The partial infilling with estuarine sediments within the dissected terraces resulted in producing flat valley floors characteristic of the box-shaped valleys (Hesp and Shepherd, 1978). The phase of estuarine deposition was succeeded by fluvial deposition and floodplain aggradation. The change between the estuarine facies and floodplain successions was a direct result of the progressive infilling of the estuary during the Holocene sea level high stand (Figs 1 and 2; Shepherd, 1987).

Holocene sedimentary successions

Coastal Dunes

Following the culmination of the Holocene marine transgressionm sand which 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-north-westerly 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 (Fig. 1). The Manawatu coast is particularly suited for aeolian transport of fine sand and the mobilisation of large parabolic and transgressive dunes. The reasons include:

- A good 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).
- Dominant onshore wave-approach, which is maintained on the Manawatu coast by the dominant west-north-west winds direction, moves sediment directly into the nearshore zone and eventually to the beach face. However, north of the Wanganui River the westerly swell encounters the coast at an oblique angle and also produces a southward longshore drift transporting sediment southward and also providing a sediment supply to the Manawatu coast. Thus, the Manawatu coast is a coast that can be classified as a constructional coast with a positive sediment budget.
- Favourable onshore wind and wave regime. The Manawatu coast is dominated by onshore west-north-west winds with winds recorded at Ohakea Air Force Base being sufficiently strong to initiate sand transport approximately 33% of the time (Shepherd, 1987, Muckersie and Shepherd, 1989).
- A modally dissipative beach profile dominated by fine-grained sand and wide gently sloping foreshores resulting in a large surface area to be exposed to drying out and the potential for aeolian transport during low tide.

The large-scale transgressive and parabolic dunes are aligned almost exactly parallel with the WNW winds and migrated inland between the Wanganui and Manawatu Rivers over the inner margins of the relatively flat Holocene marine plain to reach the floodplains of the Manawatu (Figs). The dune field is also present further south into the

Otaki-Te Hora area but is much more subdued and more restricted in its 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 running parallel with the coast (Fig. 1).

Pedological studies by Cowie (1963) identified three dune building phases. More recent research based on radiocarbon and luminescence dating from deposits associated with dunes support the pedological findings of Cowie (1963) that dune migration occurred in phases, but indicate that the phases were less distinct than originally proposed (Figs; Shepherd, 1987; Muckersie and Shepherd, 1995; Hawke and McConchie, 2006).

Foxton Phase (6500 – 1600 years ago)

The Foxton Phase of Holocene dune migration is the most extensive and oldest, 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 that the initiation of the Foxton Phase dunes was *c*. 4,000 years ago and continued until *c*. 2,000 years ago. This was based on the degree of soil development. 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. Additional evidence indicating an early to mid Holocen age for the initial deposition of the Foxton Phase dunes is the absence of Taupo Pumice (dated at *c*. 1,850±10 yr BP; Foggatt and Lowe, 1990, Shepherd, 1987).

Recent research has refined the time of active dune migration of the Foxton Phase showing that aeolian deposition was initiated soon after the culmination of the most recent PMT *c*. 6,500 and continued until *c*. 1,600 years ago (Shepherd, 1987; Muckersie and Shepherd, 1995; Shepherd and Price, 1990; Hawke and McConchie, 2005). During 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 of foredunes in the back beach environment (Gibb, 1986; Shepherd, 1987). At this time the Manawatu coast was approximately 4 km landwards of its current position (Muckersie and Shepherd, 1995; Hawke and McConchie, 2006).

It has been postulated that the initial Foxton Phase of active dune migration may have ceased near the coastline around 4,500 years ago but dunes continued to migrate inland and by 2,300 years ago had prograded over the Himatangi anticline and reached the floodplain near the junction of the Manawatu and Oroua Rivers. As the dunes migrated over the floodplain they left in their wake a thin aeolian veneer and reached their maximum extent *c*. 16 km from the coastline and stabilised *c*. 1,600 years ago (Shepherd, 1987; Muckersie and Shepherd, 1995). This is consistent with radiocarbon ages of 5,290 and 4,730 reported by Fleming (1972). Muckersie and Shepherd (1995) have also show that there was a possible hiatus of active dune formation at the coast that lasted until *c*. 3,500 years ago while the already mobilised dunes continued to migrate inland. Following the hiatus a second phase of dune formation occurred at the coast from 3,500 years ago and also lasted until *c*. 1,600 years ago during which time the Taupo Pumice would have been incorporated into the active dunes (Muckersie and Shepherd, 1995). Accordingly, two major phases of transgressive dune initiation at the coast are observed within the Foxton phase. One commenced at the coast *c*. 6,500 and the second started *c*. 3500 years ago with both operating simultaneous *c*. 3,500 to 1,600 years ago at different distances from the coast (Fig. 4; Shepherd, 1987; Muckersie and Shepherd, 1995).

Based on sedimentological analysis and eight OSL ages obtained on the Foxton Phase of dune in the Otaki-Te Horo area, Hawke and McConchie (2006) named the earlier advancement of the Foxton Phase as the Swamp Road dunes. The distinction is based on the more poorly sorted sediments with a higher magnetic content than the Foxton phase dunes. However, this is a local characteristic and is not the case for all the earlier phase Foxton dunes. The reason for the difference in sedimentary characteristic in the Otaki-Te Horo area is related to the higher energy source associated with the gravel beaches and mouth and higher energy environment associated with the Otaki River (Hawke and McConchie, 2006). Based on the OSL ages Hawke and McConchie (2006) concluded that the Swamp Road dunes and the latter Foxton Phase dunes are of a similar age (*c*. 4,000 years in the Otaki-Te Horo area), but that the Swamp road represented 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 to 2980±1070 and are consistent with the age of landward Foxton Dunes in the region of the Manawatu River (Muckersie and Shepherd, 1995; Hawke and McConchie, 2006).

<u>Motuiti Phase (1000 – 500 years ago)</u>

Dunes and sand plains of the Motuiti Phase extend up to 11 km from the 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). The dunes are characterised by having 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 was originally estimated to have been initiated < 1000 years ago. This age determination is based on Motuiti Phase dunes incorporating Taupo Pumice and overlaying Maori occupation material. The Taupo Pumice accumulated in river valleys and on beaches following the Taupo eruption of A.D. 200 and was subsequently incorporated into the aeolian phase (Cowie, 1963). This age of < 1000 years for the Motuiti Phase dunes is supported by a radiocarbon age determination of an *in situ* tutu (*Coriaria* spp.) stump rooted in soil of 855±50 yrs (Cowie, 1963). The tutu stump is rooted in the underling (preceding) Foxton Phase and is overlain by 22 m of the Motuiti Phase and based on its stratigraphic position. Cowie (1963) suggested that the tutu was killed by advancing sand (Fig. 4). Similar chronology for the Motuiti Phase dunes was also reported from the Otaki-Te Horo area where an OSL age of 670±140 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).

<u>Waitarere Phase (<120 years)</u>

Waitarere Phase dunes form a coastal belt of active and stabilised dunes between 0.5 to 4 km inland (Fig. 1; Cowie, 1963). These dunes are characterised by a lack of soil development apart from a slight darkening of the top few centimetres at the surface by organic matter and sands are relatively unweathered. This 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 and the accumulation of sand along beaches and foredune (Cowie, 1963). Vegetated dunes are extensively covered with *Scirpus nodosus* (pingao), *Spinifex hirsutus* (Spinifex) and *Ammophila arenarua* (marram grass). However there are large areas that are occupied by active foredune blowouts and parabolic dunes. Since progradation continues today, and blowouts and parabolic dunes continue to develop despite significant human intervention in the form of planting, dune re-shaping and fertilising it may be that early human activity merely aided the natural processes and the development of the most recent dune phase. Supporting evidence that active dune migration was dominantly a natural process is provided by an OSL age of 470±130 years for active dune migration from the Otaki-Te Horo area (Hawke and McConchie, 2006) which is earlier than European occupation.

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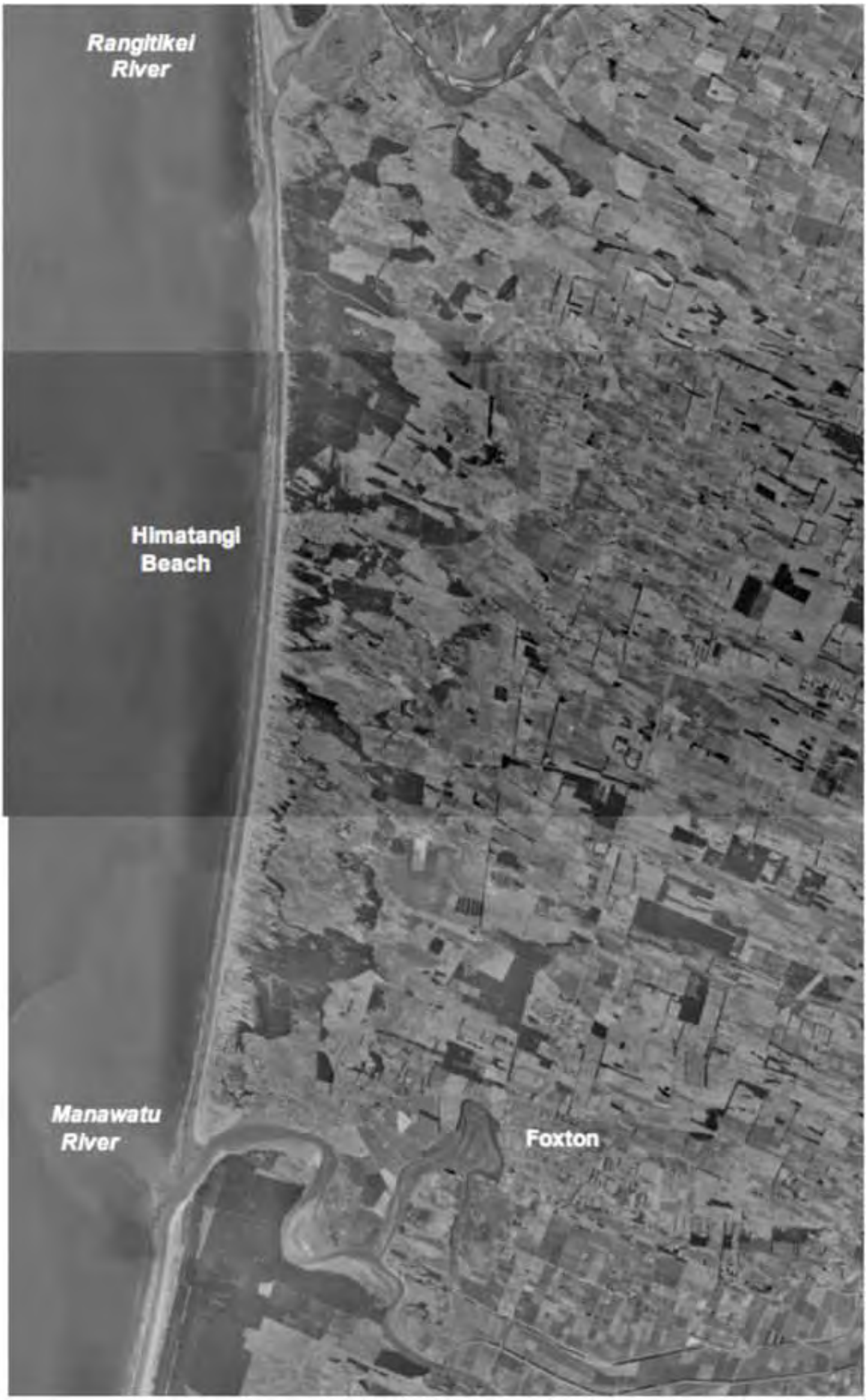
Rangitikei
River
Himatangi
Beach
Manawatu
River
Foxton