Kapiti Coast coastal hazard assessment

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Executive summary

Introduction

The Kapiti District Council contracted Coastal System Limited to provide coastal erosion hazard assessments for the Kapiti Coast, and in particular to define coastal erosion hazard distance (CEHD) lines corresponding to predicted coastal erosion over 50 years (Shand, 2008a & b), and subsequently 100 years (Shand, 2012). Potential coastal hazards other than erosion were excluded from the analysis.

The methodology used to define the CEHDs, which were renamed coastal erosion prediction distance (CEPD) lines in the 2012 report, was described as being simultaneously standard and novel. In general, the methodology follows suggested best practice for determining coastal setback lines (*viz*. Healy and Dean, 2000; Ramsay *et al.*, 2012). However, modifications are made to the methodology, and assumptions are made, that in combination indicate that the results are unfit for their intended purpose. Further, a comparison between the predicted shoreline trends using standard methodology and the observed shoreline trends indicates that the standard methodology is not appropriate, and assumed trajectories of forcing processes do not agree with observed trajectories. Therefore, an alternative approach should be used.

This report considers the Holocene evolution of the Kapiti Coast and resulting beach characteristics, evaluates the Coastal System Limited methodology and assumptions, and suggests an alternative methodology

Kapiti Coast background

Geomorphology

The Kapiti Coast between the Waikawa Stream in the north, and the Fisherman's Table Restaurant, Paekakariki, in the south, is largely an extension of the sand country that forms the coastal plains of the Manawatu (Wright, 1988). The Holocene coastal plain consists mostly of dune sequences enclosing peat swamps that lie seaward of an assumed interglacial highstand seacliff formed after sea level reached approximately the present level 7-7500 years ago (Hawke and McConchie, 2006; Gibb, 2012). The width of the Holocene coastal plain varies along the coast, starting around 3 km wide at Te Horo, reaching a maximum width of 4.2 km at Paraparaumu Beach, and decreasing to zero at Fisherman's Table Restaurant (Figures 1 & 2).

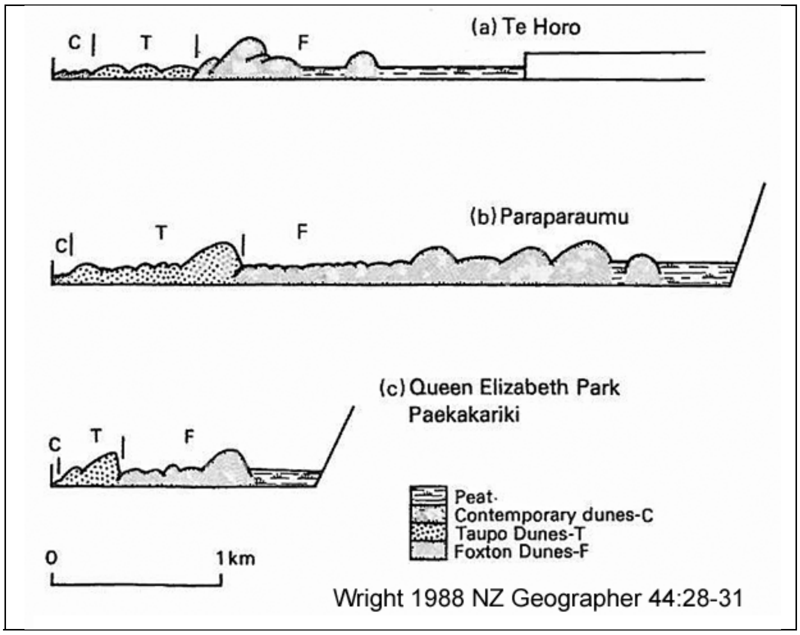


Figure 1. Schematic cross-sections of the Kapiti Coast coastal plain showing the main units identified by Wright (1988) and the varying width.

The longshore variation in shoreline position is referred to as a cuspate foreland, being generally triangular in shape and comprising of a series of shore parallel beach ridges and dunes, indicating overall offshore progradation (Craig-Smith, 2005). Although it was suggested by Wright (1988) that the cuspate foreland formed in response to wave refraction, Black and Andrews (2001) argue that due to the deep waters of the Rauoterangi

Channel, the primary mechanism is wave sheltering in the lee of Kapiti Island, and hence a reduced transport capacity. The maximum coastal plain width corresponds with the apex of the cuspate foreland (Figure 2). There is a significant longshore variation in nearshore gradient as indicated by the separation between the shoreline and the 10 m depth contour. The steepest gradient occurs between the Otaki River and Te Horo Beach, in association with mixed sand-gravel beaches, and the flattest gradient occurs between Raumati and Paekakariki (Figure 2).

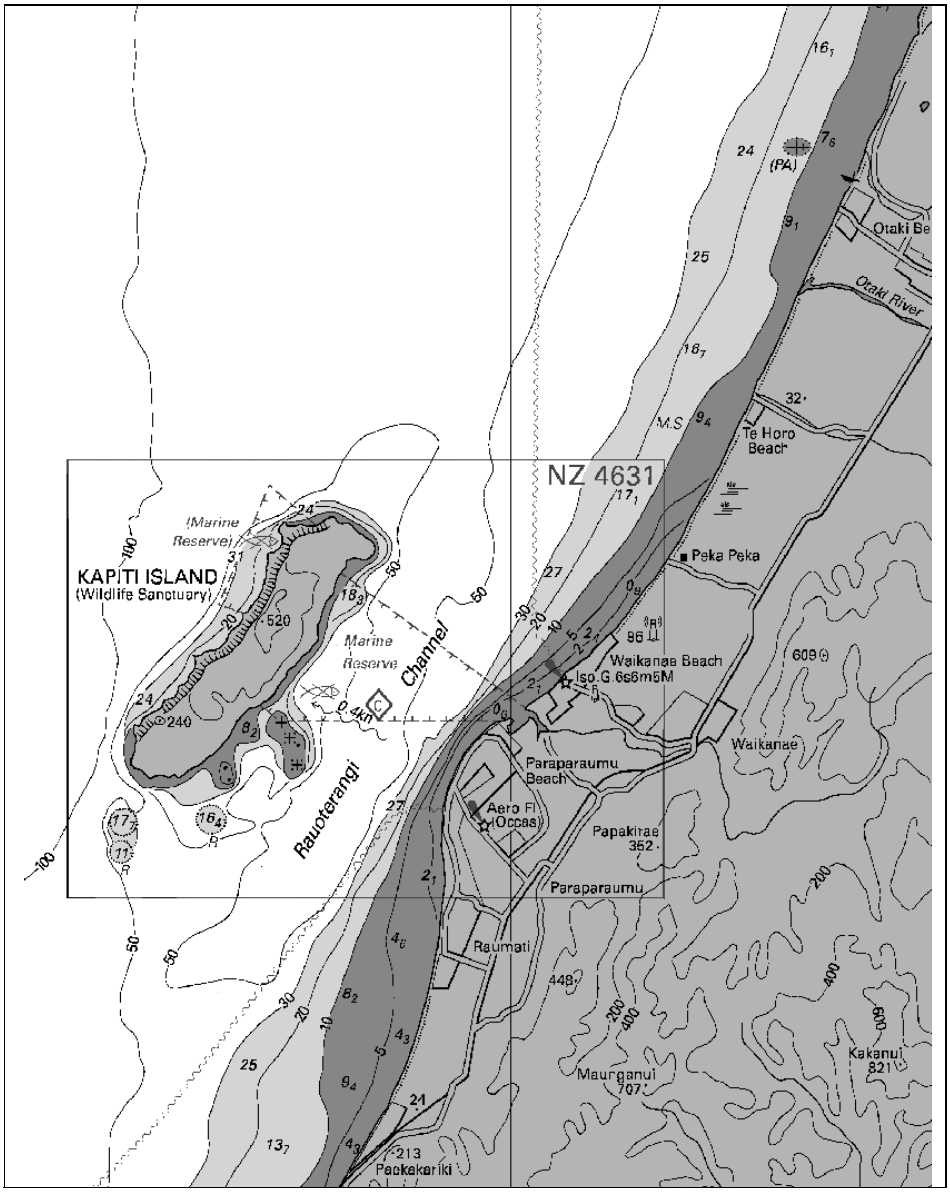


Figure 2. Section of hydrographic chart NZ 46 showing the shoreline between Waikawa Stream and Fisherman's Table Restaurant, Paekakariki. Note the cuspate foreland associated with Kapiti Island, and the varying nearshore gradient between the shoreline and 10 m depth contour

The nearshore zone narrows significantly at the apex of the cuspate foreland, with a rapid increase in water depth from 0 m to 30 m close to the shoreline (Figure 2). It is suggested that the steep slope and strong currents in the Rauoterangi Channel limit further progradation towards Kapiti Island, and hence preclude further progradation towards Kapiti Island, and hence development of a tombolo (Wright, 1988).

It has also been suggested that the proximity of deep water to the apex of the cuspate foreland results in the loss of sediment into the Rauoterangi Channel, where strong currents disperse it. However, Chiswell and Stevens (2010) demonstrate that the residual current is towards the southwest so the ridge connecting Kapiti Island to the mainland would trap sediment (Figure 2), and the maximum near bed velocities in the Channel are 0.1-0.2 m.s^{-1}, which are too low to transport sandy sediment. Further, the seabed in the channel consists primarily of rock, cobbles, and gravel with broken shell, with minor areas of mud and broken shell (Chart NZ 4631). Therefore, the Rauoterangi Channel is unlikely to be a major sediment sink for the sands transported south along the coast. It is more likely that sediment is accumulating on the inner shelf between Raumati and Pukerua Bay, south of Paekakariki (Figure 2), following the sediment transport pathway proposed by Gibb (1978) in Figure 8 of his report.

Holocene development

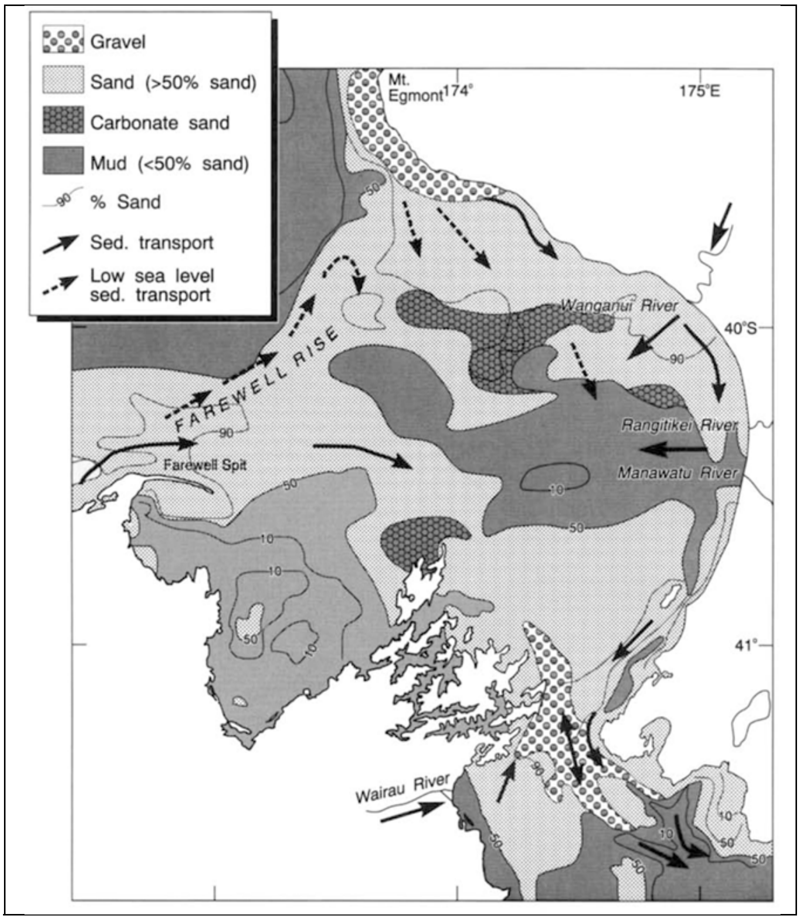


Figure 3. Summary of continental shelf sediment types between Farewell Rise and Cook Strait. (Lewis et al., 1994). Also shown are inferred sediment pathways for interglacial (solid arrows) and glacial (low sea level) conditions (dashed arrows).

Multiple sources of sediment for the Kapiti coastline have been identified. Gibb (1978) suggested that the sediment was derived from three main source regions (Figure 3) summarised below, with estimates of the present day bedload sediment discharge from Griffiths and Glasby (1985):

1. From the catchments of the Wanganui (70 $kt.y^{-1}$) and Rangitikei (40 $kt.y^{-1}$), and Manawatu (60 $kt.y^{-1}$) Rivers;
2. Smaller rivers draining the Tararua Ranges, including the Otaki River (20 $kt.y^{-1}$); and
3. Erosion of volcaniclastic deposits around Mt Taranaki/Egmont (viz Cowie *et al*., 2009).

It is also evident that small volumes of sediment are derived from the Te Paripari cliffs south of Paekakariki (Adkin, 1951), although this source is now restricted by the construction of State Highway 1 (Gibb and Depledge, 1980).

Beaches around the northern and eastern North Island coast also have derived a significant proportion of their total sediment volume from onshore movement of sand during sea level rise (*viz*. Schofield, 1970), and this process appears to be ongoing (*viz.* Bear *et al*., 2009). Wright (1988) suggests that some of the sands along the Kapiti coast represent sediment deposited on the continental shelf during previous glacials and moved onshore in response to sea level rise (marine *bulldozing* effect).

However, analysis of the sediment textural characteristics suggests the contribution from offshore is relatively small. Firstly the longshore distributions of grain size and sorting indicate a predominantly southwards movement along the shoreline from Taranaki to Paraparaumu Beach. Textural and compositional characteristics also suggest that there is a weak northwards movement from Paekakariki to Papaparaumu Beach (Gibb, 1978; Gibb and Depledge, 1980; Wright, 1989; Kasper-Zubillaga, *et al*., 2007). Secondly, the compositional characteristics of the sands between Otaki and Raumati indicate that the sediment is immature, reflecting a strong fluvial component with little modification by marine processes, and closely linked to sands found between Foxton and Wanganui predominantly derived from the Whanganui, Whangaehu, Rangitikei and Manawatu Rivers, and Kaikakokopu Stream (Kasper-Zubillaga *et al*., 2007). There is some evidence that the same sediment sources contributed to Farewell Spit, and some sediment derived from the South Island is present. This is observation is inconsistent

with the interpretation of glacial and interglacial sediment pathways of Lewis *et al* (1994) shown in Figure 3. Finally, the offshore sediment characteristics (Figure 3) indicate that the seabed is mud dominated, so there are limited sand resources directly offshore from most of the Kapiti Coast, except for Raumati to Paekakariki.

Based on 14 months of visual observations of wave conditions and the estimated volume of longshore sediment transport from Williams (1988), the present day gross mass longshore transport is of the order 80-240 $kt.y^{-1}$. This is comparable to the estimated net total mass bedload discharge from the major rivers identified as sediment sources above. It is likely that the main sediment sink is progradation of the cuspate foreland, both seaward and vertically due to inland movement of sand dunes.

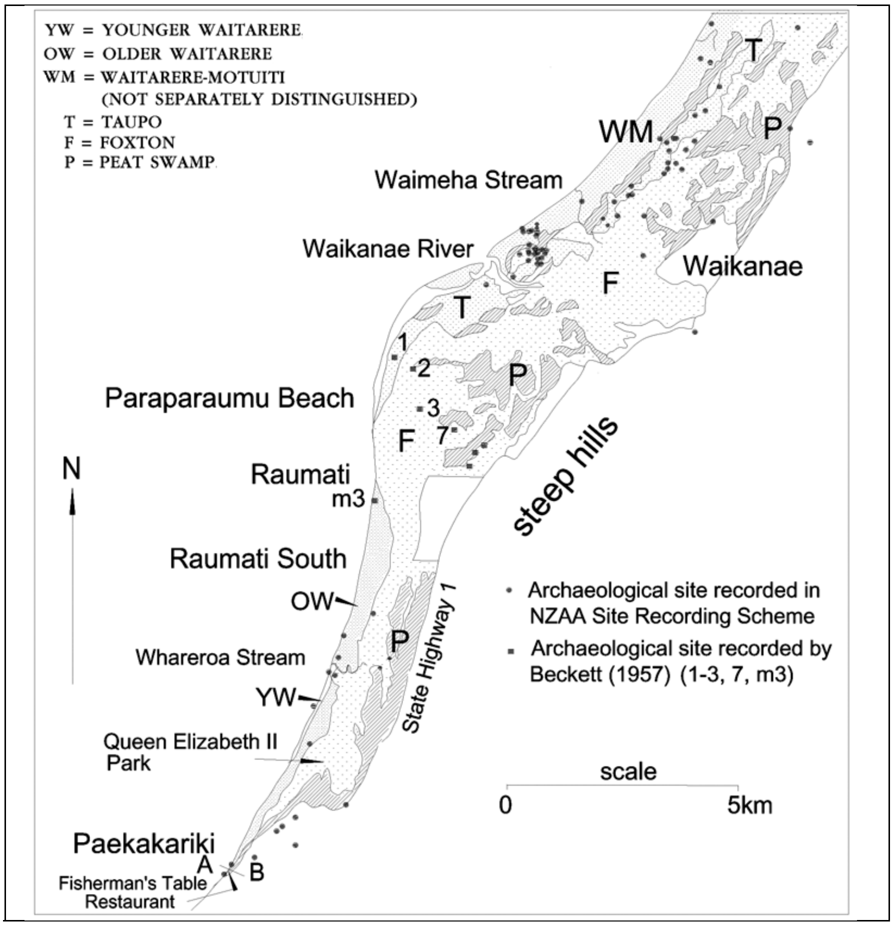


Figure 4. Sketch map of the cuspate foreland showing the main Holocene dune deposits and peat identified by McFadgen (1997). State Highway 1 approximately follows the position of the interglacial highstand seacliff.

Various studies have investigated the dune sequences of the Kapiti coast, with McFadgen (1997) providing a useful summary (Figure 4). Key dune sequences have been identified, initially based on geomorphology and soil development and subsequently by dating using ^{14}C, optically stimulated luminesence (OSL), and tephrochronology (Muckersie and Shepherd, 1995; McFadgen, 1997; Hesp, 2001; Hawke and McConchie, 2006; Clement *et al*., 2010), and these include (Figure 4):

1. *Koputaroa dunes* generally located landward of the interglacial highstand seacliff and dated at 9,000-12,700 BP. They are attributed to deposition of sand blown from braided riverbeds. Further north, an older sequence of Koputaroa dunes has also been linked to a marine source when sea level was 40-50 m below present.
2. *Swamp Road dunes* that appear restricted to the Otaki-Te Horo area, and the most landward dunes formed after sea level reached approximately the present level around 7,500 BP. These dunes are dated at 2,390-5,460 BP, and stratigraphically are considered to have formed between 4,000-4,400 BP from a marine source (as are all the younger dunes), with a fluvial input from the Otaki River.
3. *Foxton dunes* are a part of an extensive region of dunes associated with a rapid progradation of the Manawatu coastal plain between 6,500 BP and 1,600 BP. Their formation has been attributed to the onshore movement of sediment from the continental shelf associated with sea level rise. Two phases of Foxton dune development in the Manawatu can be recognised, and initial phase contemporaneous

with the Swamp Road dunes, and a younger phase contemporaneous with the Foxton dunes of the Kapiti Coast dating around 2,100-3,200 BP. The onset of the younger phase coincides with 1.5-3 m of uplift at Kapiti Island and a regional tsunami associated with a local earthquake, probably on the Wairau Fault, at 3,360±40 BP (Goff *et al*., 2000), suggesting this event may have destabilised the coastal dunes as is evident at Raumati South (Figure 5) in response to a 15th Century tsunami (Goff *et al*., 2007).

4. *Taupo Pumice*, while not directly forming sand dunes, is an important stratigraphic marker. During the Taupo Eruption of 1717 cal BP (Lowe *et al*., 2008), airfall lapilli and ash (tephra) covered the dunes, and larger sea rafted clasts were deposited on the beaches. In some areas of the Kapiti coast, the deposits of sea rafted pumice are extensive (Figure 4). These have been interpreted as marking the location of the shoreline at the time of the eruption (*viz*. Gibb, 1978). However, pumice clasts are easily broken down in the swash zone of a beach, so preservation requires that they are buried or transported inland (de Lange and Moon, 2007). Therefore, the Taupo pumice deposits identified in Figure 4 are mostly tsunami washover deposits into swales between existing dunes, similar to the Taupo Pumice deposit located in the Okupe Lagoon on Kapiti Island (Goff *et al*., 2000). Therefore, the Taupo Pumice cannot be considered a reliable shoreline marker as assumed by Gibb (1978).

Figure 5. Sand dunes at Raumati South that were remobilised by a tsunami in the 15th and then stabilised by vegetation (Goff *et al*., 2007).

5. *Motuiti dunes* are generally located seaward of sea rafted Taupo Pumice deposits, and contain significant quantities of Taupo Tephra. This suggests that they had formed around the time of the Taupo Eruption, and may have been destabilised by the tsunami that was associated with the eruption (Lowe and de Lange, 2000; Goff *et al*., 2000). They advanced over the top of Foxton dunes (Figure 6), and bury archaeological remains along their inland edge (McFadgen, 1997). Therefore, it is also suggested that human activities associated with Polynesian colonisation may also have destabilised the dunes (Clement *et al*., 2010). This dune sequences is dated between 150 and 1000 BP.
6. *Waitarere dunes* are the most recent sand dunes, being generally less than 120 years old. The youngest (OW in Figure 4) overlie European-introduced artefacts and plants, and are attributed to destabilisation of the foredunes by grazing and human activities (Cockayne, 1911).
7. *Te Horo Gravel Beach* is an important coastal unit between the Otaki River and southern Te Horo, although not strictly a dune sequence, and is of particular importance as a region of ecological significance (Forsyth and Beadel, 2012). Further, this coastal unit indicates that the Otaki River may disrupt the southwards longshore transport of sediment from the large rivers to the north (Hawke and

McConchie, 2006). The type of beach (Figure 6) progressively changes from a composite beach just south of Otaki River, to mixed sand and gravel beach near Sims Rd, to pure sand beach just south of Te Horo. Between the Otaki River and Te Horo gravel storm ridges back the beach. The ridges do not appear to have been dated, but stratigraphically correlate to the Motuiti and Waitarere dunes. The gravel storm ridges result in a significantly lower elevation of the coast plain than found for the rest of the Kapiti Coast, and it is notable that this is the only area within the Kapiti District where the 1855 West Wairapapa Tsunami appears to have inundated the coast (Grapes and Downes, 1997).

The extent of dune sequences varies along the coast (Figure 4), with each unit becoming less extensive, and fewer dunes ridges being evident from north to south. There is also some evidence to suggest that the southern dunes have been more disturbed by tectonic events than the northern dunes. Gibb and Depledge (1980) discuss evidence that the dunes around Paekakariki have undergone ~3 m of uplift while, the area around Raumati has undergone subsidence. Wright (1988) also suggests that the southern dunes were never as well developed as further north, primarily due to limited sediment supply.

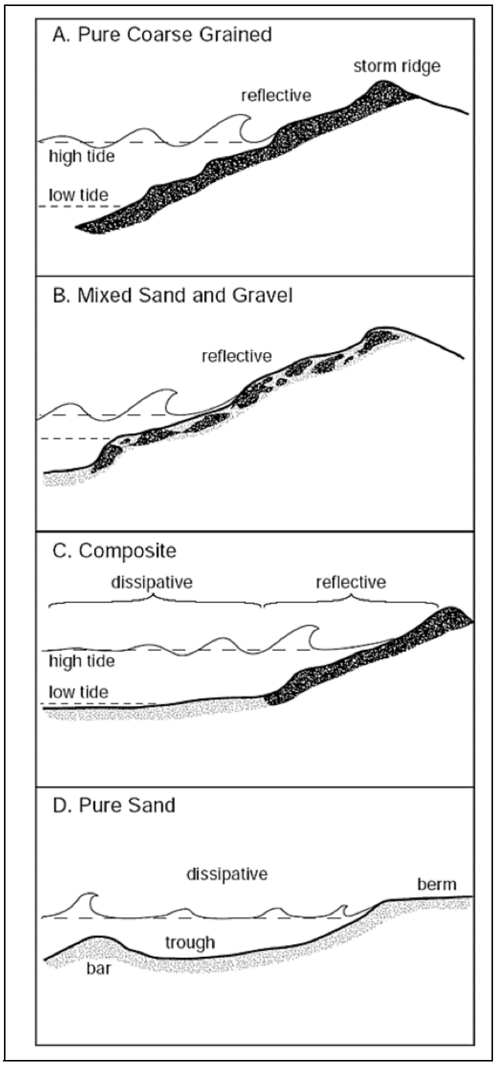


Figure 6. Types of mixed sediment beaches (Jennings and Schulmeister, 2002).

Overall, the evidence suggests that the cuspate foreland formed some time (100s to 2000 years) after the initial onshore flux of sand associated with the Holocene marine transgression. Further the growth of the foreland was primarily controlled by the southwards of sediment transport from the major river catchments to the north, leading to the asymmetrical dune development (Figure 4).

The main dune sequences are associated with phases of inland migration of sand from the coast (Hawke and McConchie, 2006), which may be initiated by either an influx of sediment to the coast (oldest Foxton dunes, and Taupo Pumice) or renewed wind erosion of previously stable dunes or other sand deposits (Koputaroa dunes, Swamp Road Dunes, Motuiti dunes, and Waitarere dunes). The most recent phases are attributed to anthropic disturbance of dune vegetation (Hawke and McConchie, 2006), although the Motuiti dune phase also coincided with at least 3 tsunami events (Goff *et al*., 2000; Goff *et al.*, 2008) as is evident at Raumati South (Figure 5).

The Waitarere dunes are most clearly linked to anthropic disruption of dune vegetation, primarily due to grazing, burning and the introduction of new flora (Cockayne, 1909, 1911; Hesp, 2001; Hilton, 2006). Cockayne (1909) reported when he surveyed the dune vegetation of the Kapiti District that "*it is not easy to say what was the typical vegetation of a fixed inland dune. The pasturing of stock, frequent burning of the vegetation, and the spread of introduced plants has, in most places, called into existence a plant-association quite foreign to primitive New Zealand*". Subsequently, Cockayne (1911) proposed the use of introduced Marram Grass (*Ammophila ar-*

enaria) as part of a strategy to stabilise the coastal dune fields around New Zealand. This was followed by the establishment of *Pinus radiata* plantations and extensive pastoral farming (Hilton, 2006).

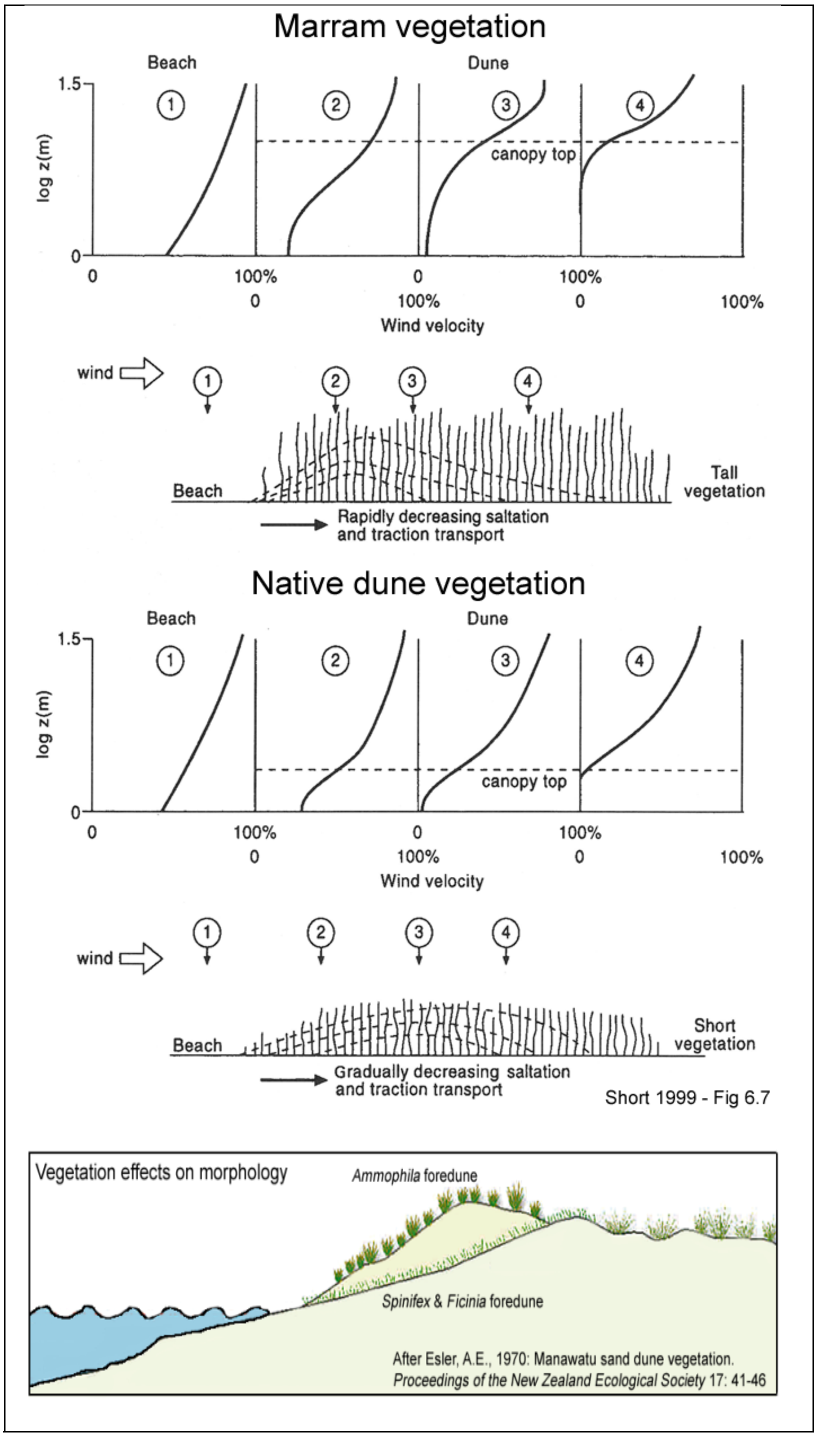


Figure 7. Effects of vegetation characteristics on foredune morphology (After Hesp, 1999).

The substitution of native dune species with Marram Grass and other introduced flora resulted in a significant change in the morphology of coastal dunes (Figure 7). Coastal dune morphological development depends primarily on vegetation density, height and cover, wind velocity and sediment supply (Hesp, 1999). Different plant species produce variations in density, height and cover. In particular, *Ammophila arenaria* produce tall dense vegetation that covers most of the surface, while the native species *Spinifex sericeus, Ficinia spiralis*, and *Asutrofestuca littoralis* produce sparse, lower vegetation with less ground coverage. These difference result in distinctly different morphologies (Figure 7).

Ammophila and associated introduced flora produced narrow high steep-faced coastal dunes to replace the lower and broader dunes that existed previously. In areas of limited sediment supply, this was associated with shoreline retreat as any given volume will occupy less horizontal space as a high steep dune. Further, during the transition from native dunes to *Ammophilia* dunes, sand was lost inland as transgressive sand sheets and parabolic dunes (Hilton *et al*., 2005). This process likely contributed to the phase of erosion between Raumati and Paekakariki reported by Gibb and Depledge (1980).

More importantly, there is growing evidence that the response of the beach to storm events differs with the morphology of the foredune. In particular, steep *Ammophilia* foredunes are more prone to scarping and collapse, while lower *Spinifex-Ficinia* foredunes are more prone to overwash that can result in accretion during storms (*Pers. Obs.*).

Dune restoration activities are now increasingly common around the New Zealand coastline, including within Kapiti District. These commonly include replanting native species to encourage the growth of foredunes, and may also involve the removal of introduced species, particularly *Ammophila*. This is resulting in the reversion of coastal morphology to pre-marram invasion conditions (Hilton *et al*., 2009).

Inlets

There are 13 inlets of varying size along the Kapiti District coastline from the Waikakariki Stream in the south, to the Waikawa Stream in the north, with the largest in terms of freshwater and sediment discharge being the Otaki and Waikanae Rivers. Most of the inlets are associated with a coastal lagoon. However, these lagoons differ from the traditional concept of coastal lagoons, which are generally tidally dominated water bodies formed as a consequence of inundation following sea level rise (Oertel, 2005). Depending on the freshwater discharge, the lagoons on the Kapiti coast are either wave or fluvially dominated, and hence behave like *hapua*, or river-mouth non-estuarine lagoons, found on the mixed sand-gravel coasts of the South Island (Hart, 2007, 2009a & b). The lagoon inlet varies in response to the freshwater discharge and volume of longshore sediment transport When the discharge is sufficiently low, the lagoons inlets become blocked and drainage occurs through the barrier as a ground water flow. At intermediate discharges, the inlet tends to migrate in the direction of longshore transport (generally southwards for inlets from Tikotu Creek northwards, and northwards for inlets south of Tikotu Creek). Finally at high discharges the barrier tends to be breached close to the freshwater channel entering the lagoon, forming a new inlet.

There is also evidence that as the shoreline has accreted, lagoons have progressively been stranded inland, forming lakes that eventually infilled with peat (Figure 4). It is possible that this has been associated with pulses of sediment transported southwards along the coast. Shand (2008b) discusses the possibility of such a sediment pulse in the late 1940s leading to extensive development of new control measures for the inlets during the 1950s.

The available evidence indicates that the natural inlets along the Kapiti Coast tended to migrate over time, and also became blocked, impeding drainage and contributing to an extensive area of swampy land between the coastal dunes and the hills (Figure 4). In order to develop the coastal plains, the swamp areas were drained, additional inlets were dug, and existing inlets were progressively modified. Since the 1920s, a range of stopbanks and training walls have been constructed around the inlets, and sediment barriers blocking the inlets have been routinely breached (Greater Wellington Regional Council, 2003; Shand, 2008b), with provision for this activity in the Regional Coastal Plan. Therefore, the present day inlets are highly modified, and limited in their ability to respond to variations in discharge and longshore sediment transport.

Relative land movements, sea level and climate effects

South of Paekakariki, three main fault zones are identified on land: Pukerua Fault, Ohariu Fault and Moonshine Fault (Gibb, 2012). The Ohariu Fault has been mapped through Kapiti District (Van Dissen and Heron, 2003), and generally follows the base of the hills flanking the coastal plains. The Pukerua Fault extends offshore at Pukerua Bay and probably links with the submarine fault systems running northwards through the Rauoterangi Channel (Nodder *et al*., 2007) on the seaward margin of the coastal plain. Further offshore, the major Wairau Fault system from the South Island is thought to continue northwards to the west of Kapiti Island. Borehole data also indicate that multiple faults disrupt the basement rock underneath the coastal plain (van Dissen and Heron, 2003).

In the Manawatu, the older deeper faults are associated with a series of anticlines that deform the surface. However, these are not evident in the Kapiti DIstrict (van Diffen and Heron, 2003). Instead, it is more likely that there is broad tilting of the blocks between the major fault zones (Gibb, 2012), down in the west and up in the east, which is consistent with the observed vertical displacements of sand dunes south of Paraparaumu Beach (Gibb and Depledge, 1980). The last identified major seismic event involved 3-4 m of vertical displacement on the Ohariu Fault around 1000-1050 cal BP. This is consistent with estimates of the onset of erosion at Paekakariki to Raumati (Gibb, 1978; Gibb and Depledge, 1980), and a tsunami event recorded at Kapiti Island (Goff *et al*., 2000).

Beavan and Litchfield (2012) reviewed long-term geological indicators and short-term continuous GPS (CPS) measurements of subsidence/uplift. For the Kapiti District they found that the geological data indicate long-term uplift of 0-1 mm.y^{-1}, that numerical models predict an upwards glacio-isostatic adjustment of 0.34 mm.y^{-1}, and that CGPS measured subsidence at 0.7-2 mm.y^{-1} (with >1 mm uncertainty).

Although there are no reliable analyses of relative sea level changes during the Holocene for the Kapiti District, Clement *et al* (2010) summarise Holocene sea level for the Manawatu region to the north, and Gibb (2012) similarly examines the evidence for the Porirua Harbour area to the south. Gibb (2012) assumes a eustatic sea level curve based on his earlier 1986 published data (Gibb, 1986), but with adjusted ^{14}C ages. The Clement *et al* (2010) combines the Gibb (1986) data with additional data, primarily from northern New Zealand) to produce a revised curve (Figure 8).

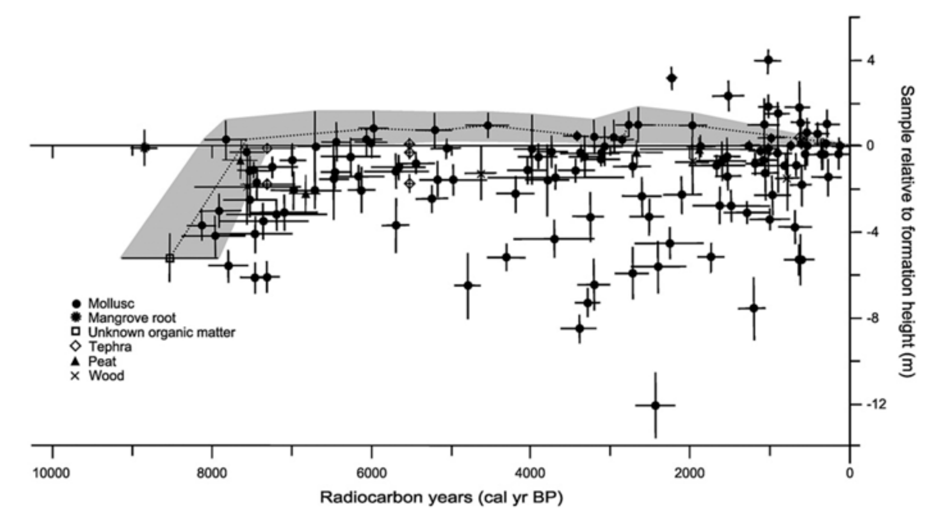


Figure 8. Revised New Zealand Holocene sea level curve (Clement *et al*., 2010).

The Gibb (2012) and Clement *et al* (2010) eustatic curves are broadly similar, but the revised curve (Figure 8) indicates sea level may have reached approximately the present position up to 1000 years earlier. Clement *et al* (2010) also indicate that the eustatic sea level was likely 0.3 m higher than indicated in Figure 8 around 7500 BP. This would make the New Zealand curve consistent with the Zone V (most of Southern Hemisphere) eustatic sea level curve of Clark and Lingle (1979), the recent assessment of the Australasian eustatic sea level curve (Lewis *et al*, 2013), and the thermosteric sea level behaviour implied by recent reconstructions of Holocene Australasian ocean heat content (Rosenthal *et al*., 2013).

Clark and Lingle (1979), and more recently Gehrels (2010), demonstrated that the concept of a single global eustatic sea level curve is misleading, and a better approach is to focus on regional sea level curves, particularly for regional planning. The key features of the regional sea level curve for the Southwest Pacific Ocean are that: the maximum sea level occurred between 7-8,000 BP; the overall trend for the last 7,000 years has been falling sea levels, consistent with the reported ocean cooling trend for this region over this time period (Rosenthal *et al*., 2013); and there have been fluctuations about the trend of the order ±0.5 m, also consistent with the fluctuations

in the ocean heat content record. Therefore, it is likely that the development of the Kapiti District coastal plain and cuspate foreland occurred during a period of fluctuating sea levels, including intervals with higher sea levels than at present.

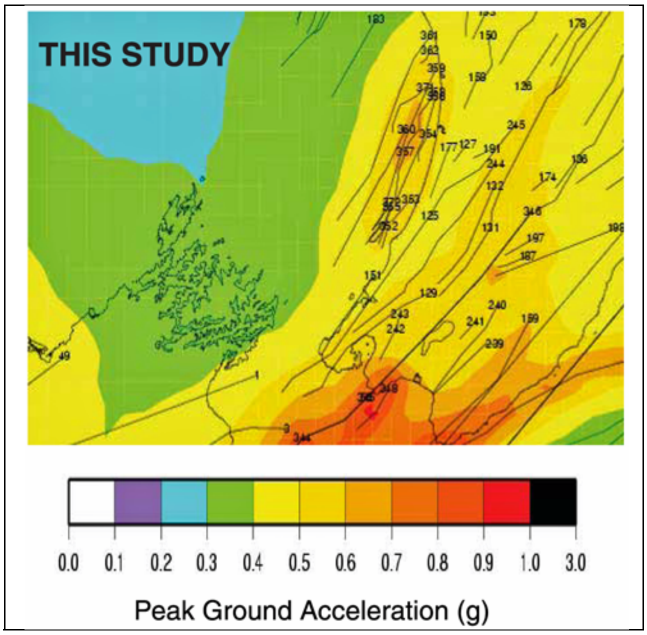


Figure 9. Distribution of faults and 475 y return period peak ground accelerations showing the influence of off-shore faults along the Kapiti-Manawatu coast (Nodder *et al.*, 2007).

Gibb (2012) also provides evidence for abrupt relative sea level changes associated with seismic events on the major faults along the west coast of the lower North Island. The mean vertical displacement during a seismic event is reported as 3.7 m, consistent with the estimated mean magnitude of M_w=6.9±0.3 for the Kapiti-Manawatu Fault System (Nodder *et al.*, 2007). The average return intervals for individual fault systems are estimated as ranging from 2,000 to >5,000 years. However, the number of fault systems present in the region results in a relatively high probability of a significant event (Figure 9).

Considering the locations of the faults in Figure 9, a seismic event causing several metres of relative sea level change is a low probability event of the order 0.02-0.05% annual probability. However, the probability of a local tsunami is higher, with annual probabilities of 0.2% for tsunami larger than 1 m based on the National Seismic Hazard Model 2010 update (Stirling *et al.*, 2012), and 0.1% for tsunami larger than 5 m based on the Goff *et al* (2000) tsunami record from Kapiti Island. The geological and geomorphic evidence indicate that either an abrupt relative sea level change, or a tsunami, can destabilise the foredunes along the Kapiti Coast, leading to parabolic dunes and transgressive sand sheets, or landward roll-over of gravel ridges. Hence, there is likely to be consequential erosion of the shoreline.

Although there is evidence for seismic events and/or tsunami triggering inland sand movement (Goff *et al*, 2008), major phases of dune migration are mostly attributed to climatic factors influencing the stability of the coastal dunes, and possibly more importantly the sediment supply (Muckersie and Shepherd, 1995; Hesp, 2001; Clement *et al.*, 2010). Allowing for variations in the underlying geology, there is a strong correlation between precipitation and sediment discharge for New Zealand catchments (Hicks *et al.*, 2011). Further, New Zealand steepland catchments appear to be particularly sensitive to environmental change at a range of time scales (Upton *et al.*, 2013). This suggests that there is likely to be a relationship between the supply of sediment to the Kapiti coast and environmental changes in the catchments draining to the coast between Cape Egmont and Paekakariki.

Grant (1981) proposed that coastal erosion around the North Island was associated with precipitation regime shifts linked to fluctuations in tropical cyclone activity. In particular, he identified an increase in storm activity that started in 1954 and continued to around 1978. Prior to the increase, there appeared to be widespread accretion around the coast, which was followed by phases of severe erosion. Increase storm activity was also associated with an increased frequency of severe floods. de Lange (2001) showed that the fluctuations in storm activity were linked to the phases of the Interdecadal Pacific Oscillation (IPO – also known as Pacific Decadal Oscillation,

or PDO, in the northern hemisphere), and they produced changes in the dominant coastal wind direction and available wave energy, which favoured periods of erosion or accretion. Proxy indicators of storm activity indicated that the fluctuations between increased and decreased storm activity had occurred for at least 5,000 years.

Although it may seem that an increased frequency of severe floods would result in a higher discharge in sediment to the coast, there is a lag in the response. Grant (1991) assessed forest disturbance within the Ruahine Range (part of the headwaters of the Manawatu River). He found that the stormy phases resulted in increased forest disturbance and mass movement, with a 2% reduction in vegetation cover and average denudation rates of 7±2 mm.y^{-1} (2-6 times the rate of tectonic uplift). The sediment that entered the channels took several decades to be transported to the coast. Grant (1991) also concluded that the fluctuations in precipitation and windiness were more significant than anthropic effects in terms of sediment discharge.

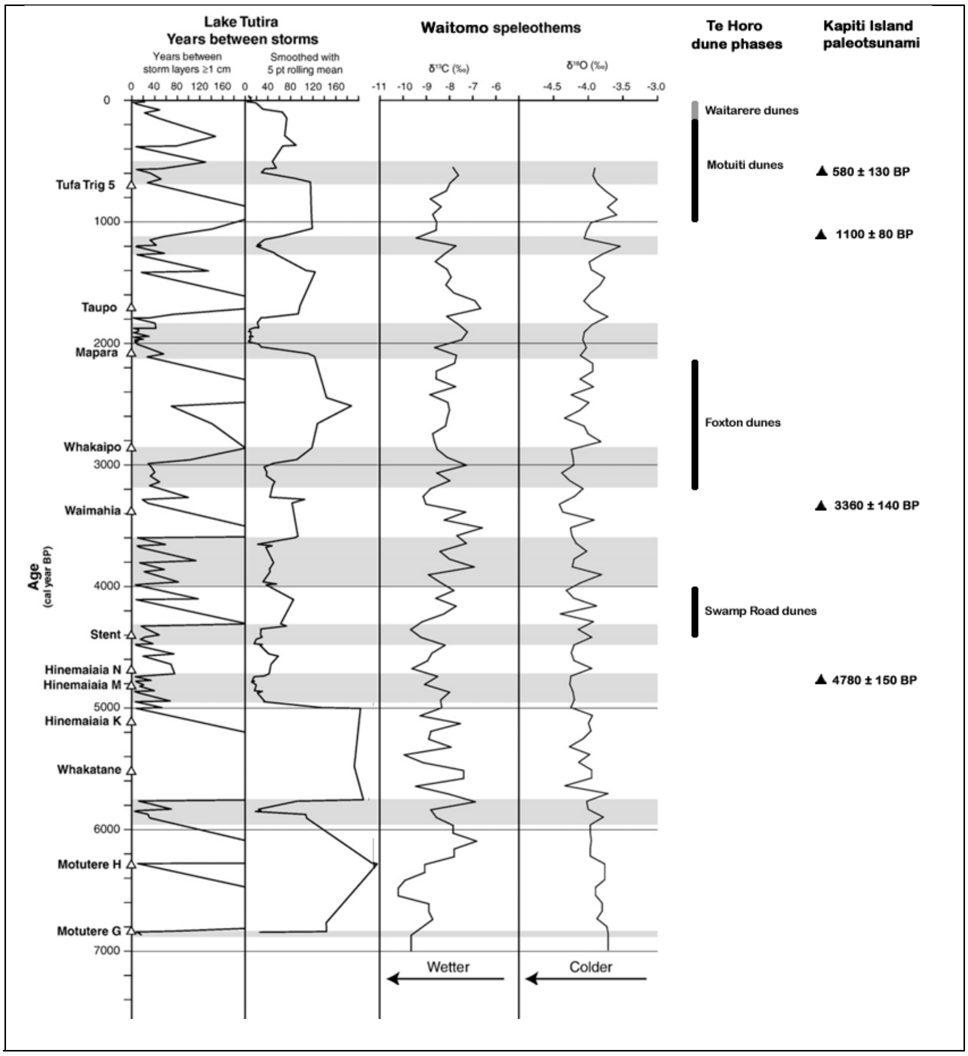


Figure 10. Comparison between storm intensity at Lake Tutira (indicated by years between storms), precipitation and temperature proxy data from Waitomo, the dune phases at Te Horo, and palaeotsunami deposits on Kapiti Island. Open triangles on the vertical axis summarise key tephra markers (After Page *et al.*, 2010; Hawke and McConchie, 2006; and Goff *et al*, 2000).

Lake Tutira, Hawkes Bay, provides a record of North Island storm activity for the last 7200 years (Page *et al.*, 2010), which was found to be a useful proxy for the discharge of sediment from the Waipaoa River catchment into Poverty Bay (Upton *et al.*, 2013). The sediment discharge from the Waipaoa River was simulated over the last 5,500 years, and found to correlate well with continental shelf sedimentation, and indicated that centennial to millennial scale precipitation fluctuations were the primary driver of changes in sedimentation rates.

Figure 10 shows the Lake Tutira storm activity measured as years between storm event deposits within the lake, climate proxy data derived from carbon (precipitation) and oxygen (temperature) isotopic ratios in speleothems from Waitomo, the dune phases preserved at Te Horo (discussed above), and the ages of palaeotsunami deposits found on Kapiti Island by Goff *et al* (2000). Page *et al* (2010) identified 25 periods of increased frequency of major storms over the last 7,200 years, of which 9 were of at least 100 years duration (shaded bands in Figure 10).

They found no relationship between storm activity and ENSO climatic variations, and speculated that storm behaviour may be influenced by the interaction of ENSO, IPO and the Southern Annular Mode (SAM). They also noted that, as is evident in Figure 10, Holocene climate for New Zealand has involved multiple periods of rapid change, particularly in terms of storm activity.

Gomez *et al* (2011) examined the Lake Tutira data in conjunction with climate proxy data from Ecuador, the Western Pacific Warm Pool, and Central Antarctica, in order to assess the combined role of ENSO and SAM climatic variations. They argue that La Niña (positive) conditions and a positive SAM both enhance rainfall and the incidence of extratropical storms and strong easterly to northeasterly winds for the eastern North Island. Hence, the storm activity record from Lake Tutira represents the relative phase of ENSO and SAM, with maximum storm activity occurring when both are positive. Although the data showed some support for this interpretation, it was also evident that the strength of the coupling between ENSO and SAM varied throughout the last 7,200 years. The variation in coupling was linked to the seasonal contrast in solar insolation, and therefore the precession component of Milankovitch Cycles, resulting in amplified responses around 5000 and 2000 BP.

Although the Kapiti District is on the west coast of the North Island, the main catchments supplying sediment to the coast (Wanganui, Rangitikei and Manawatu Rivers) all have headwaters in ranges that are affected by the same weather systems as Lake Tutira. Therefore, a similar pattern of storm activity related sediment discharge can be expected for the Kapiti District. Comparison between the Lake Tutira storm activity data and the dune phases at Te Horo (Figure 10) show that the periods of dune instability all follow periods of increased storm activity. However, not all periods of increased storm activity are associated with dune migration, and the climate proxy data (Waitomo speleothems) does not show any systematic relationship with the dune phases.

In contrast, the onset of every dune phase occurs around the same time as a major local tsunami event recorded at Kapiti Island (Figure 10). Therefore, it appears more probable that the destabilisation of coastal dunes was associated with tsunami inundation as suggested by Goff *et al* (2008), than as a direct consequence of climatic variations.

There are no published records of geological indicators of the movement of the shoreline over the last 7,000 to 8,000 years. Although the seaward margin of the dune phases and sea rafted Taupo Pumice have been suggested as shoreline indicators, these cannot be considered reliable particularly the Taupo Pumice, which probably represents an overwash deposit and not a beach deposit. Based on the available survey data (Gibb, 1978; Shand, 2008a and b), there is evidence of decadal scale pulses of sediment arriving from the river catchments. The pulses of sediment are most likely related to precipitation and windiness variations at decadal or longer scales (*viz*. Grant, 1981). Therefore, the rate of sediment supply to the Kapiti District is probably affected by variations in storm activity. However, the available evidence indicates that storm activity over the Holocene is not systematically correlated with climatic forcing. Hence, climate change is not a direct driver of sediment supply for the Kapiti Coast.

Implications for coastal erosion hazard

The Kapiti coast can be subdivided into 4 regions based on geomorphology, sediment supply, and the key processes determining coastal erosion hazard. These regions are:

1. The sandy northern flank of the cuspate foreland and northern sandy beaches between Paraparaumu Beach and Waikawa. The sediment budget is clearly positive, resulting in accretion throughout the

Holocene at an average rate of 0.4-0.6 m.y^{-1}. Accretion is continuing at present (Shand, 2008a), most likely due to bedload sediment discharge from the major river catchments to the north. However, there is some coastal erosion occurring as decadal scale cut and fill (Gibb, 1978), and possible pulses of sediment moving along the coast (Shand, 2008a). The beach systems display predominantly dissipative to rhythmic bar and trough intermediate beach states.

2. The mixed-sand gravel coast between the Otaki River and Te Horo, with associated gravel storm ridges and limited sand dune development. The sediment budget is clearly positive and appears to be primarily derived from the Otaki River, with the finer sand from further north largely bypassing (Hawke and McConchie, 2006). This area has accreted at ~0.5 m.y^{-1} over the Holocene, and is still accreting (Shand 2008a). The beach becomes progressively sandier towards the south, changing from a composite beach at Otaki River to a mixed sand gravel beach by Te Horo.

Figure 11. McLean's whare, amongst toetoe, marram and flax, on the waterfront at Paraparaumu Beach, corner of Howell Road and Marine Parade, 1914 (National Library of NZ).

3. The sandy southern flank of the cuspate foreland between Raumati and Paekakariki. Although this region has accreted over the Holocene, including since the Taupo Eruption according to Gibb (1978), the rate decreases to essentially zero at Fisherman's Table Restaurant. Gibb (1978) identified two regions of long-term erosion that primarily correspond to areas of urban development, particularly the construction of dwellings on the early 1900s foredune. The first subdivisions occurred in 1906 around Raumati and 1907 at Paekakariki, coincident with the establishment of *Ammophila* for dune stabilisation (Figure 11). Gibb (1978) also indicated that accretion had occurred in the central region occupied by Queen Elizabeth II Park. Shand (2008a) identifies this entire zone as undergoing erosion, and suggests that the 1880 and 1958 shorelines determined by Gibb (1978) were incorrect. The beaches are predominantly dissipative to longshore bar and trough beach states.

4. The inlets along the coast are strongly affected by freshwater discharge, and therefore are considered as a separate coastal type. Although there is some tidal influence for most of the inlets, overall they behave more like non-estuarine river mouth lagoons than estuarine lagoons. The frequency and magnitude of flood events, the volume of bedload sediment transport, and the magnitude of longshore sediment transport affect their behaviour. Some of the inlets were created to facilitate drainage of the coastal swamps, most have been modified for at least 80 years as part of flood management works, and the Otaki and Waikanae Rivers have been used as sediment sources, particularly for gravel (Williams, 2011).

It is evident that a single methodological approach to assessing coastal erosion hazard is inappropriate. Shand (2008a & b) accordingly used separate analyses for the open coast and inlets. However, given the differences in

prehistoric and historic behaviour for the 4 zones identified, the open coast should not be treated as one type of morphodynamic system.

Coastal System Ltd methodology

The 2008 reports distinguished between coastal areas directly affected by stream and river discharge to the coast (Part 2: Inlets) and the rest (Part 1: Open coast). Different methodologies were used to determine the CEPD for the two types of coastal areas, and these are discussed separately below.

Open coast erosion

The basic equation used is a version of the "standard" methodology used for Coastal Hazard Zonation (Healy and Dean, 2000; Ramsay *et al*., 2012), with no scaling factors evident in the relationship as expressed in Equation 1 (page 11, Shand, 2008a), and the inclusion of a combined uncertainty term. This equation is

$$CEPD = LT + ST + SLR + DS + CU$$

where

1. *CEPD* = Coastal erosion prediction distance (changed from CEHD = coastal erosion hazard distance terminology between the 2008 and 2012 reports).
2. *LT* = Longer-term historic change based on 50 year time period from cadastral maps and aerial photographs. Ideally, this should be a minimum of 60 years to allow for the fluctuations due to climatic oscillations such as IPO and SAM;
3. *ST* = Shorter-term historic fluctuation. From the discussion in Shand (2008a) this was to be derived from statistical analysis of historical data, but it practice it wasn't. For a stable or accreting coast, this is probably the most important term;
4. *SLR* = Shoreline retreat associated with sea-level rise induced by global warming. Strictly this should be due to the affect of an acceleration in sea level rise, as historic sea level rise is already incorporated into the *LT* term;
5. *DS* = Dune stability. This accounts for the scarp retreat to a stable slope after an erosion event, assuming that the previous terms are predicting the location of the base of the slope and infrastructure of concern is located at the top of the slope;
6. *CU* = Combined uncertainty. This is defined by Shand (2008a) as the error associated with the previous four terms in the equation, and any other precautionary measures that result from assumptions made in the analysis.

The methodology used by Shand (2008a; 2012) to determine each of these terms is considered in more detail below:

LT – long-term trend derivation and uncertainty

The long-term trends were derived from aerial photographs, and pre-digitised shorelines determined by NWASCO predominantly from aerial photographs and unspecified cadastral maps. It was noted that a systematic error resulting from using vegetation lines as shoreline indicators in aerial photographs, and reported high tide shoreline on the cadastral maps produced an over-estimate of shoreline erosion rates. These shoreline indicators may be several to tens of metres apart depending on beach state.

A landward reference point was used to define 68 locations (there are 59 C locations in the spreadsheet provided), and the distance between the shoreline and reference point measured in GIS (presumably but not stated?) from the geo-rectified aerial photographs and NWASCO plotted shorelines. It is assumed that the geo-rectification results in a location error of ±3 m, with a further error in estimating the shoreline position of ±3 m (for both aerial photos and NWASCO shoreline data?).

For each location about 9 measurements were made from aerial photographs, and 1-2 from cadastral map shorelines. These should have different uncertainties.

The long-term trend was determined by Ordinary Least Squares (OLS) regression analysis. Three different trends were determined:

1. Entire record – 1870s to 2007
2. Early period – 1870s to early 1950s
3. Late period – 1940s to 2007

These dates are not exact because the survey coverage varies along the coast. The early period was assumed to be unaffected by coastal management. This is not correct. The dunes were affected by grazing and burning resulting in extensive vegetation loss and destabilisation. Following the Sand Drift Act (Introduced 1903, enacted 1908) the dunes were planted in Marram Grass, which significantly altered their shape and behaviour (Hilton, 2006). It can also be argued that land-use changes and flood protection works in the catchments have affected sediment yield over the entire record.

Although other NZ studies have identified long-term patterns of shoreline fluctuation and Grant (1981) identified these patterns for the Kapiti Coast (as confirmed by Dr Shepherd in the "review" attached to Shand (2008a)), this study has decided to treat "non-linear" trends using break-point analysis without any constraints on the minimum trend duration (Figure 3 Shand, 2008a). This approach has a significant effect on the long-term trend. In particular, it replaces a long-term (~100 y) trend with one based on a few decades. In figures 3A and 3C (Shand, 2008a), it changes an accretionary trend into erosion, which is misleading. In Figures 3B and 3D (Shand, 2008a), the magnitude of the trend is altered significantly.

It is claimed that, apart from the sites in Figure 3 (Shand, 2008a), the late period trend was *qualitatively* similar to the trend over the entire record.

The report also compares the early period trend with the late period trend. This is important for assessing the impact of coastal structures, particularly since the analysis later considered scenarios where the structures are removed or fail. The difficulty is that the early period analysis typically compares 1-3 cadastral survey data points with 1-2 aerial points. Since there is a difference in the shoreline definition between the two types of data that biases the trend, the inferred trends are meaningless. This is acknowledged in the report as "*Given that these rates may be exaggerated by the inclusion of tide-based shorelines from cadastral maps, and affected by lack of intermediate data-points, the pre-urban shoreline appears to have been relatively stable*." (page 20 2008-2). Therefore, it seems to assume that in the critical area where structures now exist, the long-term rate prior to construction is "stable"?

Overall, the long-term trend is derived from the late period trend, except for those sites with seawalls or a "recent trend change" (Figure 4B). Using the data supplied, the ratio of the late period trend to the early period trend varies from -2 to 814, with an average of 25. This is a very large variation, and also suggests that overall

there has been an increased rate of accretion over time. Those sites with a recent trend change use the short-term trend determined by the weighted linear model (strictly appears to be a truncated linear model using selected recent data points).

Sites with seawalls are assumed to have no long-term trend while seawalls are present. However, the report notes that there has been accretion at some seawall sites (in one case the seawall is completely buried now - site C12.50). What happens when seawalls are removed?

So for the calculated rates of shoreline movement there are:

1. Trends determined by OLS for the 1940s to 2007 (late period) – a trend over a maximum period of 67 years, which is barely long enough to span the 50-70 year fluctuations in NZ shorelines identified by other studies.
2. Trends determined by "weighted" OLS for the 1990s to 2007 (non-linear sites) – which is really a short-term trend.
3. "Stable" areas with no trend due to the presence of sea walls.

Then, if the trend is positive (coast is accreting) it is set to zero. Also, if the short-term weighted OLS trend is for faster erosion than the late period trend, it is used as a long-term trend.

Hence, a coast that the data shows to be predominantly accreting is transformed into either "stable" or eroding.

The uncertainty in the LT factor is determined as follows:

1. The assumed geo-rectification (±3 m) and shoreline detection errors (±3 m) are combined to give an assumed error of ±4.2 m.
2. The longshore variation of the "error" in the OLS regression for the late period data was assessed and an estimated 95% upper percentile was used to represent the entire coast. It is not clear exactly which error is referred to, but it appears to have been the Standard Error of Estimate, which is the standard deviation of the residuals.
3. Other factors that affect the uncertainty are discussed but then ignored.

The error that should be relevant to the LT factor is the uncertainty in the gradient of the OLS trend (ie. the uncertainty of b in Equation 2). This measures how much faster or slower the shoreline can be moving relative to the estimated rate. The report states that this was ignored because "the weighting procedure, together with the variance reduction measures of setting positive rates to zero and the selection of the maximum longshore rate, were found to be adequate" (page 26).

The report also states that the ±3 m shoreline detection error was found empirically to produce a ±3.7 m error in the "rates of change" over a 50-year prediction period. Apart from the inconsistent units, it is not evident how this was calculated and why? However, this number is taken to be the LT uncertainty for the entire coast. Further, it is assumed that the only uncertainty to take into consideration is -3.7 m.

Hence, setting all accreting coastal sites to zero, and then applying an LT uncertainty of -3.7 m transformed the entire Kapiti coastline transformed into an erosional zone (-0.074 m/y *cf.* a long term trend of 0.4-0.6 $m.y^{-1}$ for most of the coastline).

ST – Short-term shoreline fluctuation and uncertainty

The short-term shoreline fluctuation accounts for the cut and fill associated with storm events. Analysis of this fluctuation can be complicated for several reasons:

1. The erosion phase (cut) is considerably faster than the recovery phase (fill); typically being hours compared to days to decades for the complete return of eroded sediment volume. Usually, up to 80% of the recover occurs within days to a few weeks if most of the eroded sediment is transported offshore into the offshore bar.
2. If sediment is transported onshore by wave overwash, there may not be a significant recovery phase. This is particularly important for coarser sediments (mixed sand-gravel, and gravel beaches).
3. Storms may occur in clusters, so that the beach may not fully recover before a subsequent erosive event occurs. Studies around the NZ coast have identified that there has been decadal-scale fluctuations in storm frequency and magnitude, which means that a coast can show an erosive trend for several years to decades, followed by an accretionary phase.

Analysis of the short-term fluctuations requires a time-series data-set that captures the short duration erosion events, as well as the longer duration recovery phases and the decadal-scale effects of storm clustering. It is evident that the aerial photograph and cadastral survey records used for the 2008 study were not suitable for characterising the short-term trend. Some beach profile data were available, but were not utilised. However, from the description of the profile data sets (footnote page 27, Shand 2008a), the time series data are not suitable for characterising the short-term fluctuation.

In Shand (2008a), it is assumed that the short-term fluctuations are represented by the residuals between the measured shoreline location and the trend line. This is not a reasonable interpretation for several reasons:

1. The shoreline position was recorded using two different approaches: cadastral survey of high water mark or toe of the foredune; and vegetation line determined from aerial photographs. These would correspond to different shoreline positions, even if taken at the same time, and would appear as residuals from the trend. However, late period trends should involve only one type of measurement;
2. The vegetation lines are not likely to represent the average shoreline position (assumed by this approach). As noted in the report, the vegetation line retreats during erosion, and takes time to return to the original position after shoreline recovery. Therefore, the vegetation line is biased towards an eroded shoreline.
3. The residual approach assumes that the rate of erosion/accretion is constant over time (linear trend). It is likely that this is not the case, as the sediment supply and driving processes are not constant.
4. The errors in geo-rectification and shoreline position determination appear to be of a similar magnitude to the calculated standard error of estimates. Therefore, a component of the residual is likely to represent the measurement errors.

Therefore, the variations represented by the residuals probably do not represent the short-term cut and fill fluctuations. It is also of concern that the residuals appear to be the error term considered for the uncertainty of the *LT* factor, and therefore have been incorporated in the summation more than once?

Interestingly, the estimated *ST* does not appear to correlate well with measured short-term fluctuations along the Kapiti coast, and Appendix C discusses this. In my opinion, the discrepancy arises because the methodology used (3xSEE) does not reflect the true short-term fluctuations of the shoreline.

The uncertainty was derived from the measurement errors using an undefined empirical method that gave an uncertainty of ±2.6 m.

For the CEPD summation, only negative values for *ST* and the uncertainty were considered. Again, for an accreting coast experiencing cut and fill, this approach will generate an erosional hazard.

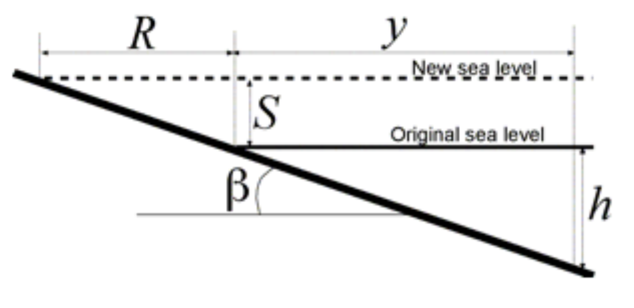
There was also an assumption of a 5 m erosional uncertainty if the existing seawalls are maintained, due to vertical scour in front of the structure. It is not clear how the vertical scour translates into horizontal erosion in the presence of a stabilised shoreline?

SLR – Impact of sea level rise determination and uncertainty

This factor is included to account for sea level rise anticipated as a consequence of global warming. Since the *LT* factor already includes the effects of historic relative sea level changes and is extrapolated into the future, the *SLR* factor should strictly be based on the changed rates of sea level rise or fall over the period of interest. This was not done, so the *SLR* factors calculated will be biased too high.

Appendix D of the report assessed shoreline response models to sea level rise. It confuses the Bruun Rule with later variations of it, such as the Dubois model (presented as Equation D1) and mostly discusses estimates of the closure depth. This is largely irrelevant as most studies have found that the most effective estimate of nearshore slope is based on the surf zone, or the steepest parts of the submarine nearshore zone. Equation D1 attempts to approximate this by including the height of the sub-aerial berm or foredune.

Overall, Bruun type approaches have been found to be unsuitable for predicting shoreline response to sea level rise. In particular, using the approach to hindcast the effect of historic sea level rise has been found to normally overpredict erosion (Note the method will only predict erosion, and clearly does not work for an accreting coast).



$$R = \frac{S}{\tan\beta} = \frac{yS}{h}$$

Figure 12. Definition sketch for the mathematical formulation of the Bruun Rule for the shoreline retreat due to sea level rise initially proposed by Bruun (1962).

It is suggested that the Komar et al equation is a better alternative. This was intended to predict the extent of storm cut during a single event, and the methodology developed by Paul Komar and his students differs from Equation D2 presented in the report. Equation D2 is the original formulation of the Bruun Rule (Bruun, 1962).

In line with the earlier 2008 report, the 2012 report states that a commonly used approach for assessing the response to coastal, known as the Bruun Rule, is inappropriate for the Kapiti Coast because many of the assumptions of the rule are violated. The mathematical concept behind the Bruun Rule is very simple for anyone who can remember basic trigonometry, being based on similar right-angled triangles (Figure 12).

Essentially, the Bruun Rule states that the shoreline retreat is equal to the ratio of the sea level rise to the slope of the shoreline. There are different ways of defining the slope ranging from the slope of the continental slope, to the slope of the beach face or offshore bar. The ratio indicates that a low slope will result in more retreat that a steep slope, so by selecting the "right" slope it is possible to get a desired retreat.

It is also evident from the ratio, that a higher sea level estimate will also produce more shoreline retreat. Again there is the opportunity to choose the "right" sea level rise to get a desired answer. The key difference between the 2008 and 2012 assessments for the Kapiti Coast, is the use of much higher estimated sea level rise with no consideration of the probability of these occurring.

Finally, it should be obvious that this Rule cannot predict shoreline accretion, as *R* will always be positive. The problem is that for the New Zealand beaches where the rule has been tested, in the long-term *R* should be negative because the beaches are accreting. From the evidence presented in the Kapiti Coast assessment, *R* should be negative for the Kapiti Coast, and the authors are correct to state that the Bruun Rule is inappropriate.

Instead they have used a method proposed by Komar *et al* (1999) and applied to the Oregon coast of the USA. Dr Jeremy Gibb has also applied this method to coastal hazards in New Zealand. This method assumes that the key factor driving dune erosion is saturation of the sand at the base of the dune, and it is developed to predict shoreline erosion due to storm events (see sketch).

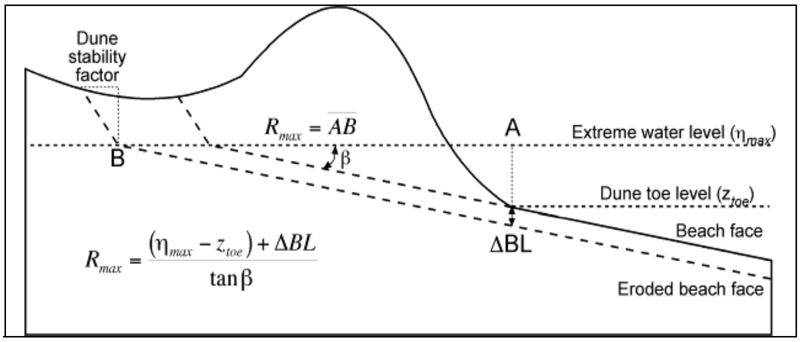


Figure 13. Definition sketch for the foredune erosion model in response to storm events proposed by Komar *et al* (1999)

This method does not include a sea level rise term at all. It is based on the extreme water level relative to the elevation of the dune toe, and the beach lowering during the storm. It is not intended to predict the effects of long-term sea level rise. Consequently the authors of the Kapiti Coast study modified the Komar *et al* equation by replacing the numerator term with the sea level rise (Equation 3 in the report). Therefore, instead of using the equation as defined, they used the ratio of the sea level rise to the slope of the beach. I have already discussed this ratio – it is known as the Bruun Rule!

The method used depends on the nearshore slope, which was taken to be the inter-tidal beach slope, and the predicted change in sea level. For the Kapiti coast, nearshore slope was estimated for 22 sites that do not seem to coincide with the 68 coastal hazard calculation sites (details hidden in the database?).

The nearshore slopes estimated varied between 0.8° and 6°, although most were around 1-2°. Using Equation 3, the predicted sea level rise is multiplied by 9.5 to 71.6, with most locations having a multiplier of 28.6-57.2. These relatively high multipliers reflect the generally dissipative to intermediate beach state along the Kapiti coast. Note that based on the measured shoreline response to the historic sea level rise of order 17 cm/century assumed in the report, the multipliers should be predominantly negative (-247 for the average accretion rate of 0.42 m/y).

The other component is the predicted sea level change. Both the 2008 and 2012 reports are based on various projections of future sea level derived from economic scenarios used to estimate future radiative forcing, and hence future temperatures. The projections then assume that sea level responds in a predictable manner to glo-

bal temperatures. So far this has not been the case, and more than 40 years of sea level projections have not successfully the actual global sea level response.

Of concern is the lack of probability associated with the sea level projections. Although terminology such as *most likely value* is often applied to sea level projections, this is a qualitative judgement and not a statistical interpretation. The 2008 report is based on a value of 0.6 m/Century, while the 2012 report used 0.6 m/Century for the 50-year projection (0.3 m total) and 0.9 m/Century for the 100+ year projection. In my opinion, these values are excessive and improbable, and I note that the 2008 report considered the assumed sea level rise or 0.6 m/Century to be "conservative" (page 34, 2008-2), which in context reflected an expectation that it would result in an overestimate of shoreline response.

For stabilised coasts (with seawalls), it is assumed that sea level rise will not cause erosion.

The *SLR* uncertainty is based solely on the estimated error in the measurement of the nearshore slope, and was determined to be ±1.6 m. I am uncertain as to why the slope measurement was converted to an angle for this determination. The slope error was originally ±0.001 grad, and, since this the calculation effectively takes the reciprocal of the slope, the error analysis should have been based on percentage error. The *SLR* uncertainty should consider the uncertainty of the sea level projections.

DS – Dune stability factor determination and uncertainty

The *DS* factor takes into account the slope adjustments that occur after an erosion event, particularly the scarp retreat that results in an additional landward migration of the upper dune face, assuming that the erosion has scarped the frontal dunes. In relation to the Kapiti coast assessment, this scarp adjust has already been accounted for because the shoreline is based on the vegetation line (ie. landward of any scarp, after a period of time during which it is likely that the face has adjusted to a stable angle). As discussed above, the *LT* and *ST* factors are both based on the vegetation line and will already include any *DS* adjustment. Therefore, for the CEPD the *DS* term is double-dipping.

The methodology used to assess *DS* is standard, and assumes that the material falling from the top of the slope accumulates at the toe until a stable slope is achieved. The result depends on the assumed stable slope angle (34° for this analysis) and the height of the scarp. It was assumed that the future scarp height equated to the maximum dune height near the profile for sites south of Otaki, and equal to the maximum for the entire Kapiti coast for sites north of Otaki. This is only valid if the final future erosion event terminates coincident with the maximum dune height. Overall, the approach used will over-estimate *DS*, as noted in the report (page 36, 2008-2).

The uncertainty is based on the RSS measurement error for the estimated maximum dune height, and was calculated as ±2.3 m. It does not include any consideration of the uncertainty in the assumed stable slope angle.

CU – Combined uncertainty determination

There are some issues with the approach to the uncertainty as expressed in the definition of Equation 1 in the original report:

1. Some factors are time dependent (*LT* and *SLR*, which involve multiplying a factor by the time interval being considered) while others are not (*ST*, which is a fluctuation about zero, and *DS,* which is a one-off adjustment). Strictly the uncertainties of the time dependent factors will increase with time, and the others will not.

2. It is not clear why there should be additional uncertainty factors beyond those that are already incorporated into the uncertainties of *LT*, *ST*, *SLR* and *DS*.

The uncertainties derived for the LT, ST, SLR and DS factor were combined using the Root Sum Squares (RSS) approach. The report states that the *CU* factor was also included in the RSS summation, but this appears to be incorrect (it shouldn't be included). It was also stated that the 4 factors are independent. However, the LT and ST factors are highly correlated and their uncertainties were derived from the same measurement errors by unspecified empirical methods. Hence, I would not consider them to be independent.

The calculated CU factor was ±5.3 m, which was rounded up to ±6 m.

Inlet methodology

Where a stream or river discharges at the coast a tidal inlet typically forms. Different types of inlets can form depending on the balance between freshwater discharge, tidal flows and longshore sediment transport. The type of inlet is not too important for a hazard zone assessment, but the amount of inlet migration is a factor. Over time the inlet position can move along the coast, generally in the direction of longshore sediment transport, with erosion on the downdrift side and accretion on the updrift side of the inlet forming a longshore spit and tidal lagoon. There tends to be a maximum amount of lateral movement, as flood events tend to breach the longshore spit and effectively straighten the inlet. The spit may also be artificially breached to achieve the same effect.

It is argued that for the Kapiti coast, the hazards associated with tidal inlets are significantly different to those experienced on the intervening open coasts. To account for this, the open coast CEPD equation was modified by replacing the short-term fluctuation with an inlet migration factor (*IM*).

$$\begin{aligned} IEPD &= LT + IM + SLR + DS + CU \\ &= IM - (LT + SLR + DS + CU) \end{aligned}$$

The *IM* factor is not clearly defined in the report.

Coastal Systems Limited Assumptions

At each step of the determination of the CEPD, the analysis maximises the estimated future shoreline erosion. This is described as "conservative" and "precautionary". Of particular concern is that this approach ignores any mitigating factors, except for the presence of seawalls. Overall, it has the effect of exaggerating the future hazard and almost certainly has identified areas that are unlikely to experience any coastal erosion as being hazardous.

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