Kapiti Coast coastal hazard assessment

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Executive summary

The Kapiti Coast consists of a coastal plain that merges with cuspate foreland that has been accreting in the lee of Kapiti Island since sea level reached a maximum between 7,000 and 8,000 BP. The current average rate of accretion varies between 0.4-0.6 $m.y^{-1}$, which is consistent with the long-term rate over the Holocene. Despite the overall trend for accretion, some areas have experienced coastal erosion that has affected coastal properties since 1900. The areas consistently affected by erosion are located south of the Tikotu Creek (Raumati and Paekakariki).

The sediments of the Kapiti coastal plain are primarily derived from the major rivers to the north (170 $kt.y^{-1}$) and local rivers (28 $kt.y^{-1}$). The supply of sediment appears to be affected by climatic oscillations influencing precipitation and windiness, potentially resulting in a cycle of longshore sediment transport of 50-60 years duration. This cycle appears to significantly affect the migration of inlet systems along the coast.

There is no compelling evidence of any relationship between prehistoric and historic shoreline movement and sea level and climatic changes for the Kapiti Coast. There is evidence that local earthquakes producing abrupt changes in relative sea level, and tsunamis have affected the shoreline stability.

The methodology adopted by Coastal Systems Ltd (CSL) was analysed, and this report discusses the various aspects that influence the Coastal Erosion Prediction Distance (CEPD) lines produced. The major concerns with the methodology are:

1. The methodology systematically maximises the CEPD at almost every step in the process in order to produce a conservative result. Consequently, the predicted CEPD lines greatly overestimate the risk of coastal erosion for the Kapiti Coast. Hence, it is unreasonable to assume that all of the properties seaward of the CEPD will experience erosion during the prediction periods of 50 or 100 years. The available data indicate that there is in fact a low risk that the majority of properties seaward of the CEPD will be affected by coastal erosion within this time period.
2. Components of the methodology used have been recognised as inappropriate for the purpose. The methodology also did not consider the morphodynamic differences along the coast associated with changing sediment type and foredune vegetation, which influence erosion processes and hence erosion hazard.
3. A risk assessment of coastal erosion hazard should include a probabilistic analysis of the drivers and impacts related to coastal erosion. This was not done, so there are no data to quantify risk, or permit a cost-benefit analysis of any proposed management responses.

Applying the CSL methodology as a hindcast for the interval 1950-2007 demonstrated that the methodology is a very poor predictor of past coastal erosion (4% success compared to 87% assuming past trends). This does not provide confidence in the reliability of the methodology for predicting future coastal erosion. Given the identified problems, the CSL methodology cannot be used to make an assessment of the risks of coastal erosion at any point on the Kapiti Coast, and an alternative probabilistic approach should be utilised.

One alternative approach is to evaluate the sediment budget the Kapiti Coast, in order to identify areas unlikely to stop accreting, those that may start eroding in the future, and those that are in sediment deficit. At present the average accretion rate for the Kapiti Coast is of the order 1.2 $kt.y^{-1}$, which is 2 orders of magnitude smaller than the available sediment supply (~200 $kt.y^{-1}$). Therefore, it is unlikely that most of the shoreline will change to a long-term sediment deficit.

The determination of the CEPD lines should differ to account for the availability of sediment. Areas with a sediment surplus, and hence accreting, should require a CEPD primarily based on the short-term storm event erosion. This is best determined from shore profile data, which would provide the probability distribution for shoreline recession caused by storms.

Areas with an existing or potential sediment deficit should be subject to a process-based probabilistic analysis of the CEPD. An example for the Kapiti Coast based on the methodology of Ranasinghe *et al* (2012) is given in the report.

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Introduction

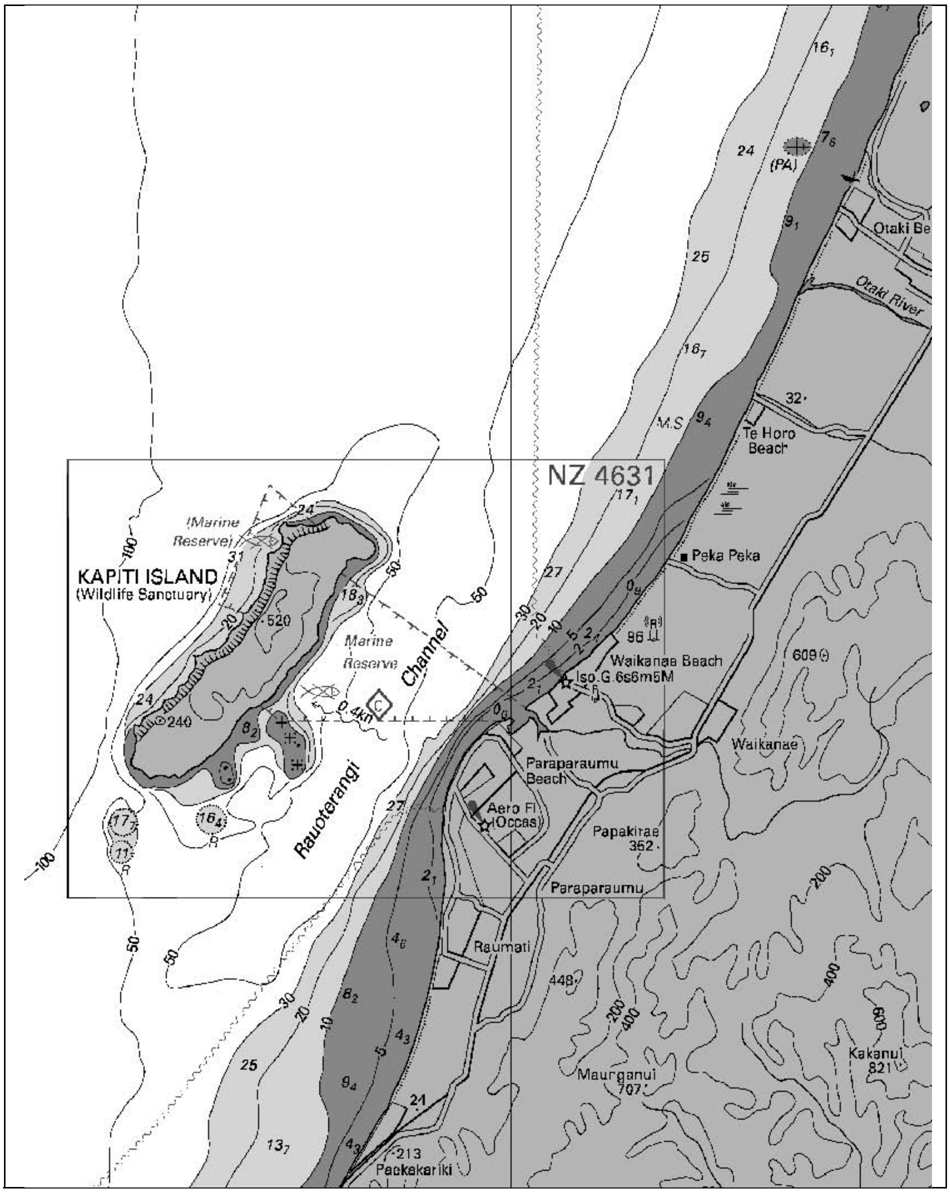


Figure 1. Section of hydrographic chart NZ 46 showing the Kapiti District shoreline between Otaki Beach and Fisherman's Table Restaurant, Paekakariki. Note the cuspate foreland associated with Kapiti Island, and the varying nearshore gradient between the shoreline and 10 m depth contour

The Kapiti Coast District Council contracted Coastal Systems Limited to provide coastal erosion hazard assessments for the Kapiti Coast (generally shown in Figure 1), and in particular to define coastal erosion hazard distance (CEHD) lines corresponding to predicted coastal erosion over 50 years (CSL, 2008a & b), and subsequently 100 years (CSL, 2012). Potential coastal hazards other than erosion were excluded from the analysis.

In general, the approach used to define the CEHDs, which were renamed coastal erosion prediction distance (CEPD) lines in the 2012 report, follows what has been best practice for determining coastal setback lines in terms of the individual components that should be considered: long-term trends; short-term fluctuations; changes in forcing processes; and characteristics or stability of coastal sediments (*viz*. Gibb, 1983; Healy and Dean, 2000; Ramsay *et al.*, 2012). However, this methodology does not consider the probabilities associated with the components, and hence does not provide a probabilistic assessment of risk, which is a requirement of risk management coastal planning frameworks (Ranasinghe *et al*., 2012).

Further, CSL (2008a) modified the methodology used to determine the individual components of the CEPD lines, and made assumptions that appear to reflect planning interpretations and not objective science, that in combination indicate that the results are unfit for their intended purpose.

Comparison between predicted shoreline trends using standard methodology and the observed shoreline trends indicates that the standard methodology is not appropriate (*viz*. List et al, 1997; Cooper & Pilkey, 2004; FitzGerald *et al*., 2008), and assumed changes of forcing processes do not agree with observations (de Lange and

Carter, 2013). It has also been recognised that better methods for assessing coastal hazards are required that do incorporate a probabilistic estimate of coastal response to sea level (*viz*. Ranasinghe *et al*, 2012). Therefore, an alternative approach should be used.

This report considers the Holocene evolution of the Kapiti Coast and resulting beach characteristics, evaluates the Coastal Systems Limited methodology and assumptions, and suggests an alternative approach to assessing the risk of coastal erosion.

Kapiti Coast background

Geomorphology

The Kapiti Coast between just north of the Waiorongomai 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 7,000-7,500 years ago (Hawke and McConchie, 2006; Gibb, 2012). The width of the Holocene coastal plain varies along the coast, being 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).

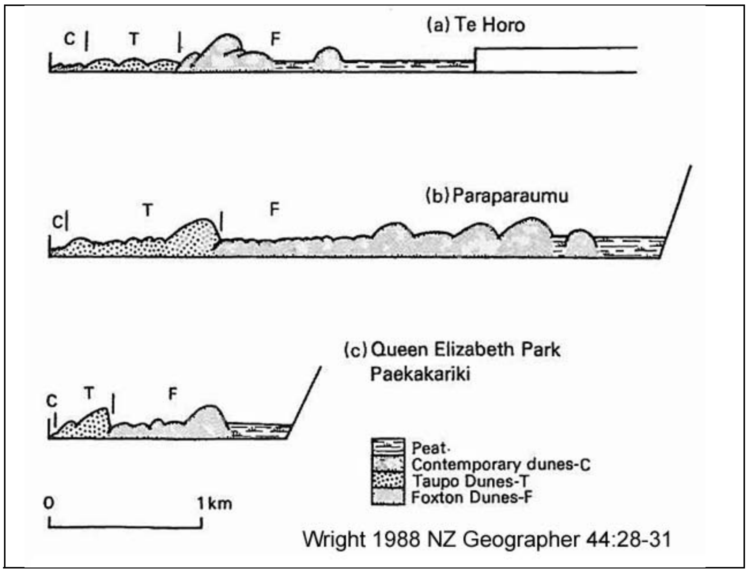


Figure 2. Schematic cross-sections of the Kapiti Coast coastal plain showing the main units identified by Wright (1988) and the varying width.

Cuspate foreland

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 1). 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 1).

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 1). 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 (Wright, 1988). 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 1), 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 1), following the sediment transport pathway proposed by Gibb (1978) (Figure 10 below).

Holocene development

Sources of sediment

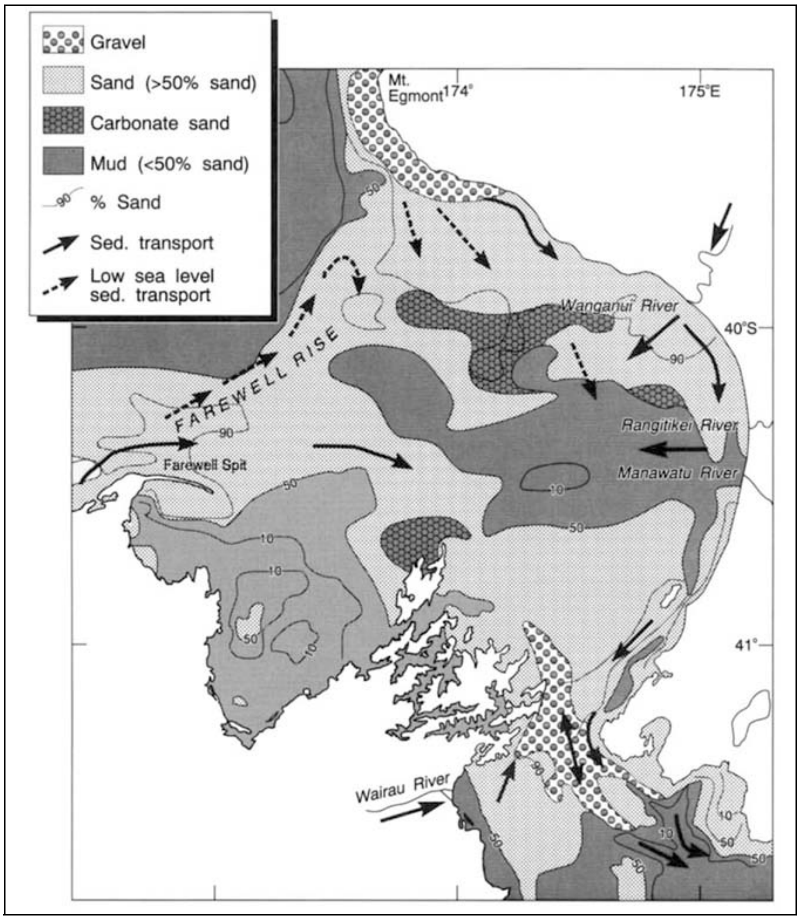


Figure 3. Summary of continental shelf sediment types between Farewell Rise and Cook Strait. (Lewis et al., 1994). Also shown are inferred sediment pathways for interglacial (solid arrows) and glacial (low sea level) conditions (dashed arrows).

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 Waikanae River (8 $kt.y^{-1}$); 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 may have been 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 Paraparaumu 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 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 and LINZ Chart NZ 4631) indicate that there is a zone of mud dominated seabed along the coast, so there are limited sand resources directly offshore from most of the Kapiti Coast, except for the shallow area between Kapiti Island and the coast between Paraparaumu Beach and 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.

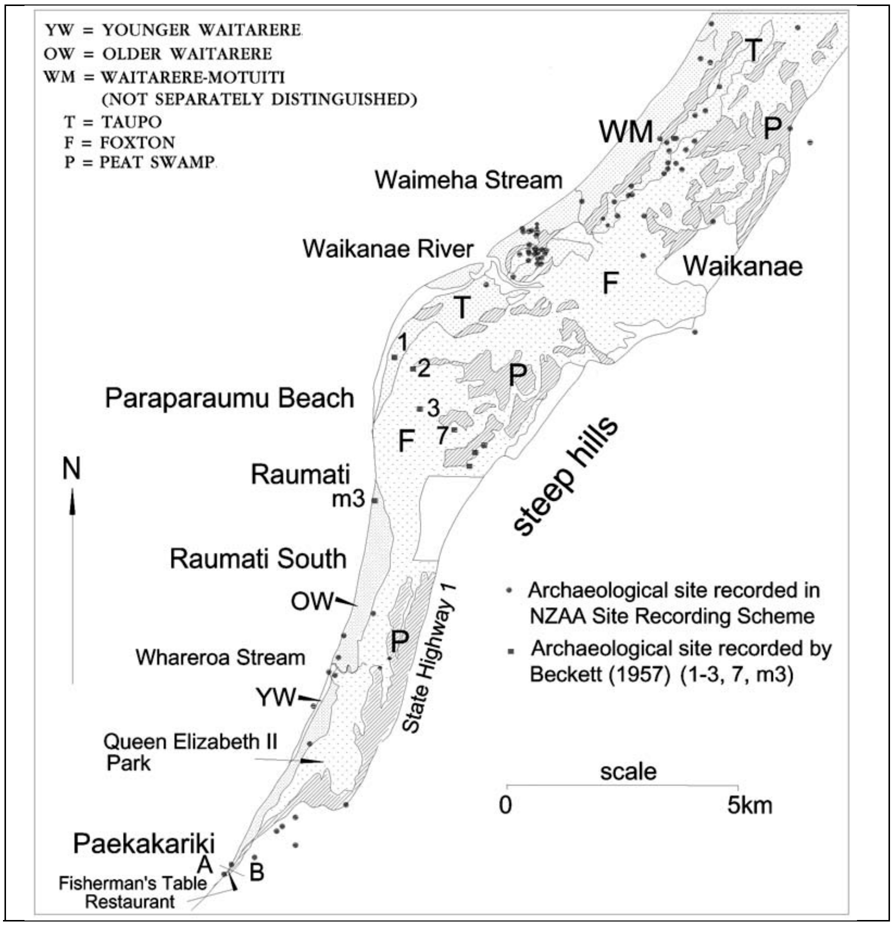


Figure 4. Sketch map of the cuspate foreland showing the main Holocene dune deposits and peat identified by McFadgen (1997). State Highway 1 approximately follows the position of the interglacial highstand seacliff.

Dune sequences

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 luminescence (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 do not appear in Figure 4. These are 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, an 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). Hence, the Taupo Pumice deposits identified in Figure 4 are mostly tsunami washover deposits formed in swales between existing dunes, similar to the Taupo Pumice

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

deposit located in the Okupe Lagoon on Kapiti Island (Goff *et al*., 2000). Thus, the Taupo Pumice cannot be considered a reliable shoreline marker as assumed by Gibb (1978).

5. *Motuiti dunes* (labelled as WM in Figure 4) 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, and bury archaeological remains along their inland edge (McFadgen, 1997). Therefore, it is suggested that human activities associated with Polynesian colonisation may also have destabilised the dunes (Clement *et al*., 2010). This dune sequence is dated between 150 and 1000 BP.
6. *Waitarere dunes* are the most recent sand dunes, being generally less than 120 years old. McFadgen (1997) separates them into Old and Young Waitarere dunes (OW and YW respectively in Figure 4) based on buried artefacts and vegetation types. The youngest dunes overlie European-introduced artefacts and plants, and are attributed to destabilisation of the foredunes by grazing and human activities (Cockayne, 1911).
7. Mixed-sediment beaches are associated with the discharge of gravel-sized sediment to the coast. The major zone of mixed sediment beaches is the *Te Horo Gravel Beach* between the Otaki River and southern Te Horo Beach, which 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). Following the classification of Jennings and Schulmeister (2002), the type of beach progressively changes from a composite beach just south of Otaki River, to mixed sand and gravel beach near Sims Rd, to predominantly sandy beach just south of Te Horo. Between Otaki River and Te Horo, gravel storm ridges form the coastal plain immediately inland from 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. A smaller extent of mixed-sediment beach occurs at the southern end of the coast at Paekakariki. This area is highly variable depending on sediment availability.

The extent of dune sequences varies along the coast (Figure 4), with each unit becoming less extensive, and fewer dunes ridges being evident progressing 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.

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 southwards sediment transport from the major river catchments to the north, leading to asymmetrical dune development (Figure 4).

Influence of dune vegetation

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).

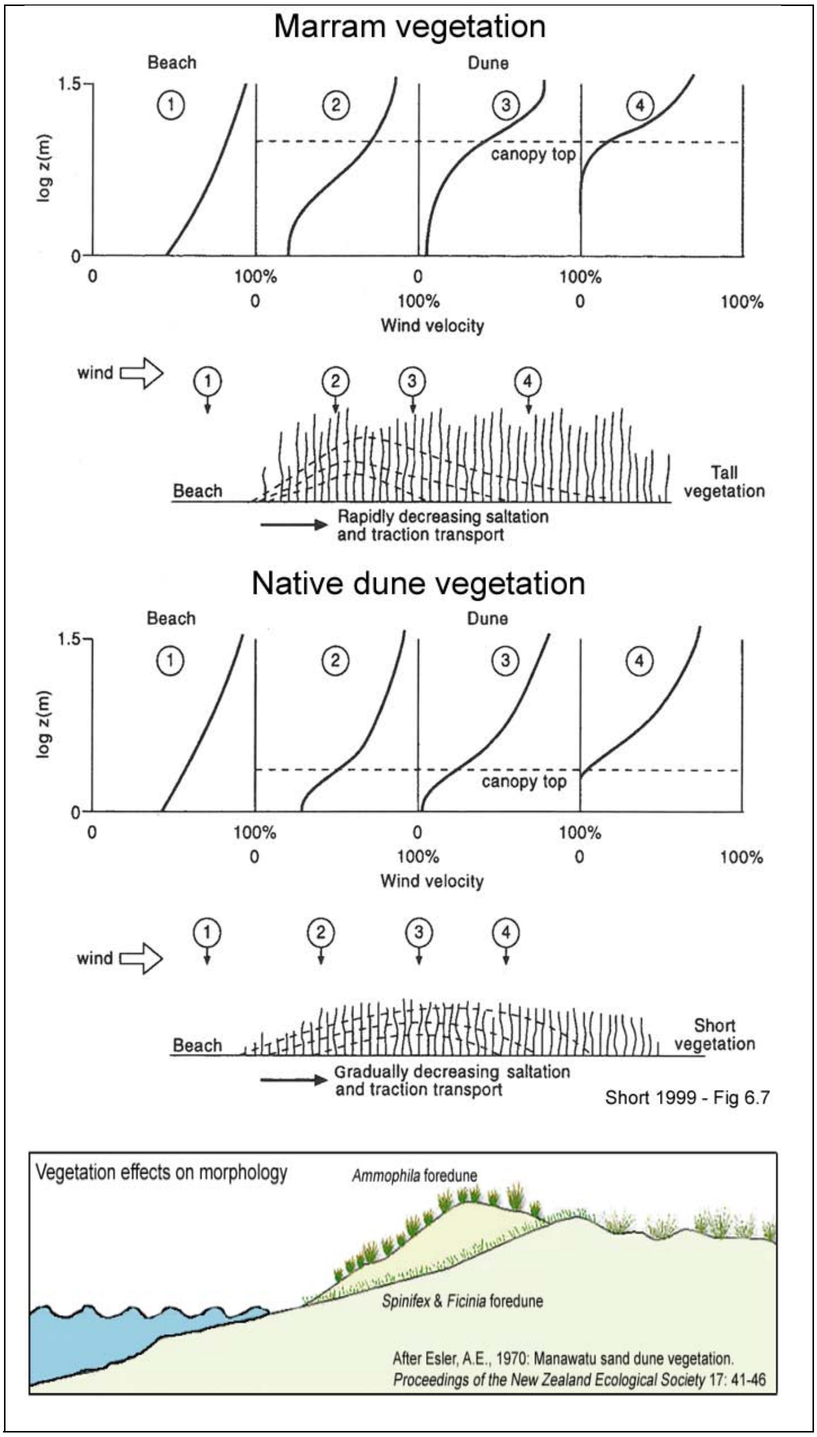


Figure 6. Effects of vegetation characteristics on foredune morphology (After Hesp, 1999).

The Waitarere dunes are 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 "*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 arenaria*) 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 then extensive pastoral farming (Hilton, 2006).

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 6). 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 differences result in distinctly different morphologies (Figure 6).

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 *Ammophila* 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 *Ammophila* 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 12 inlets of varying size along the Kapiti District coastline from the Waikakariki Stream in the south, to the Waiorongomai 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). For these systems the lagoon inlet varies in response to the freshwater discharge and volume of longshore sediment transport, with several distinct phases being recognised (Hart, 2009a):

1. When the discharge is sufficiently low, the lagoons inlets become blocked and drainage occurs through the barrier as a ground water flow.
2. 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 – CSL (2008b)).
3. Finally at high discharges the barrier tends to be breached close to the freshwater channel entering the lagoon, forming a new inlet.

The shoreline changes mapped by CSL (2008b), indicate that this pattern of behaviour occurs at inlets on the Kapiti Coast. 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. CSL (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 some of the inlets, and sediment barriers blocking the inlets have been routinely breached (Greater Wellington Regional Council, 2003; CSL, 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 Dissen 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. Clement *et al* (2010) combines the Gibb (1986) data with additional data, primarily from northern New Zealand, to produce a revised curve (Figure 7).

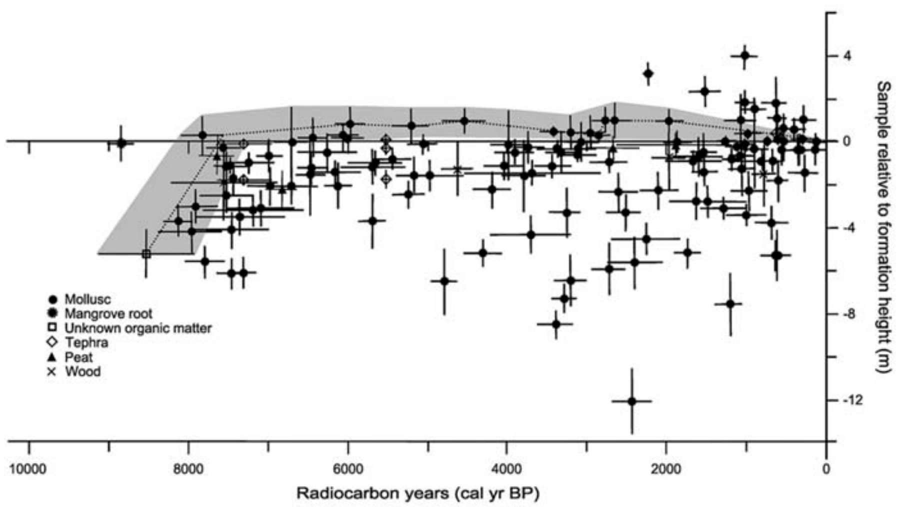


Figure 7. Revised New Zealand Holocene sea level curve (Clement *et al*., 2010).

The Gibb (2012) and Clement *et al* (2010) eustatic curves are broadly similar, but the revised curve (Figure 7) 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 7 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. The sea level rise observed at the Kapiti Coast at present is consistent with the pattern over the last 7,500 years.

Shoreline response to eustatic sea level rise

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. There is no clear relationship between regional sea level variations and the shoreline response along the Kapiti Coast; accretion has occurred regardless of whether sea level rose or fell.

Shoreline response to abrupt relative sea level rise

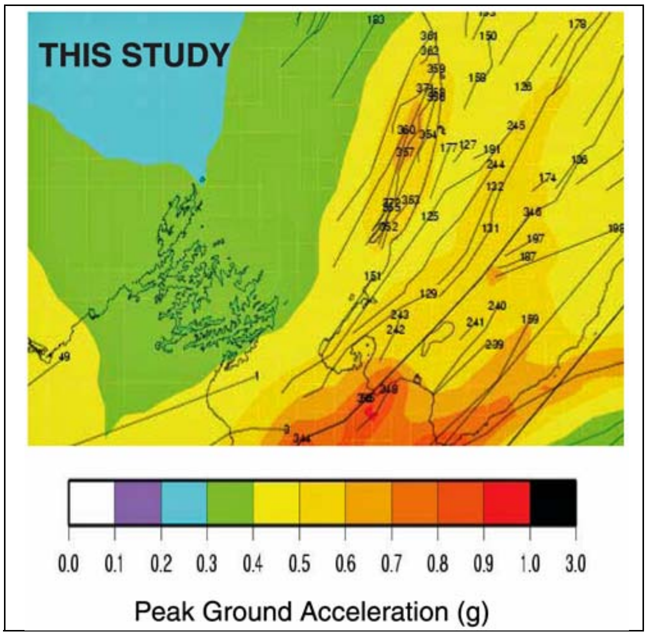


Figure 8. Distribution of faults and 475 y return period peak ground accelerations showing the influence of offshore faults along the Kapiti-Manawatu coast (Nodder *et al.*, 2007).

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 8).

Considering the locations of the faults in Figure 8, 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.

Impacts of storm activity on sediment supply

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. Increased 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 an increased frequency of severe floods results in a higher discharge in sediment to the coast, there is a lag in the response so this effect is not contemporaneous with the flood events. 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.

Impacts of climate on storm activity

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 9 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 9). They found no relationship between storm activity and ENSO (3-7 year) climatic variations, and speculated that storm behaviour may be influenced by the interaction of ENSO, IPO (50-60 year fluctuations) and the Southern Annular Mode (SAM). They also noted that, as is evident in Figure 9, 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.

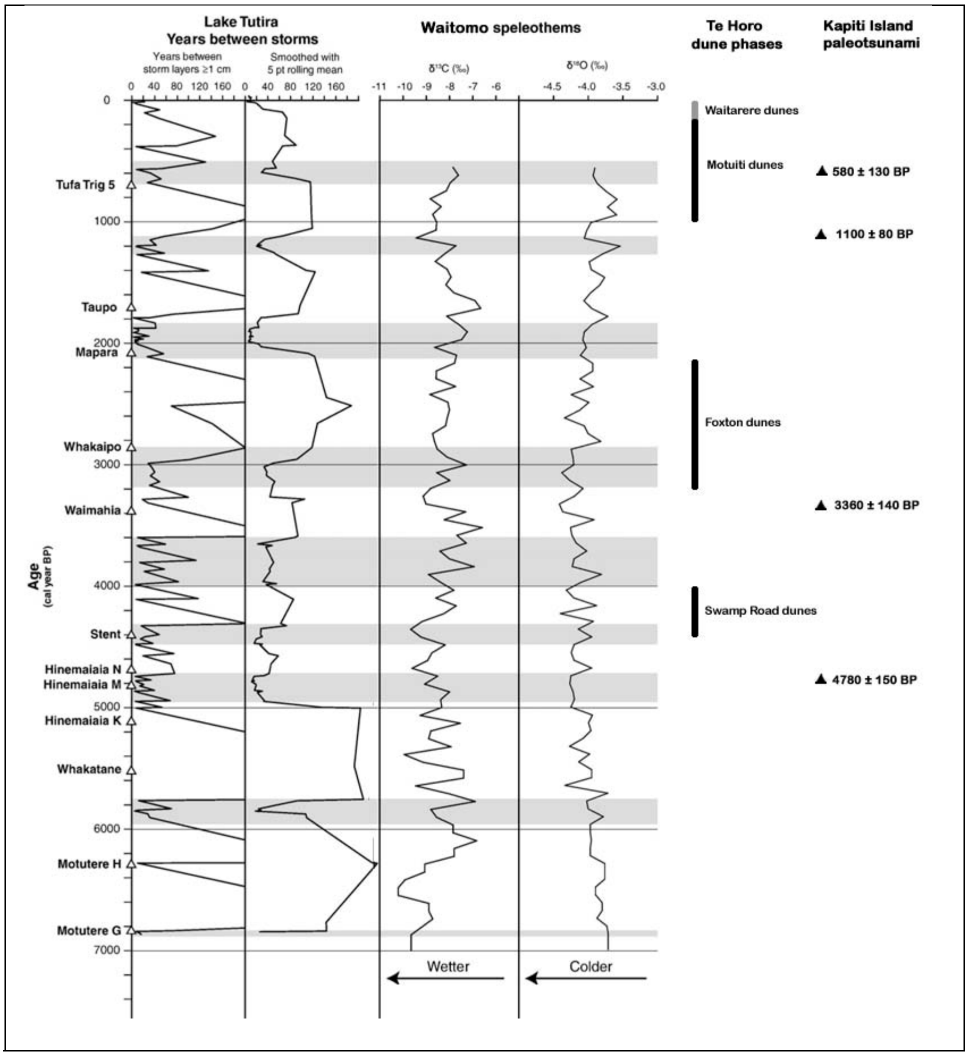


Figure 9. Comparison between storm intensity at Lake Tutira (indicated by years between storms), precipitation and temperature proxy data from Waitomo, the dune phases at Te Horo, and palaeotsunami deposits on Kapiti Island. Open triangles on the vertical axis summarise key tephra markers (After Page *et al*., 2010; Hawke and McConchie, 2006; and Goff *et al*, 2000).

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 9) 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 9). 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; CSL, 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.

Conceptual model of sediment pathways

Gibb (1978) proposed sediment transport pathways affecting the stability of the coast between Paekakariki and Paraparaumu Beach (Figure 10). The key features are a southward movement of sediment from major sources in the north, which is deflected offshore near the apex of the cuspate foreland, and a northward movement of sediment from sources south of Paekakariki. Longshore sediment transport converges between the Wharemauku Stream and Tikotu Creek, and the offshore deflection of sediment transport leads to deposition on the inner shelf between Paekakariki and Raumati.

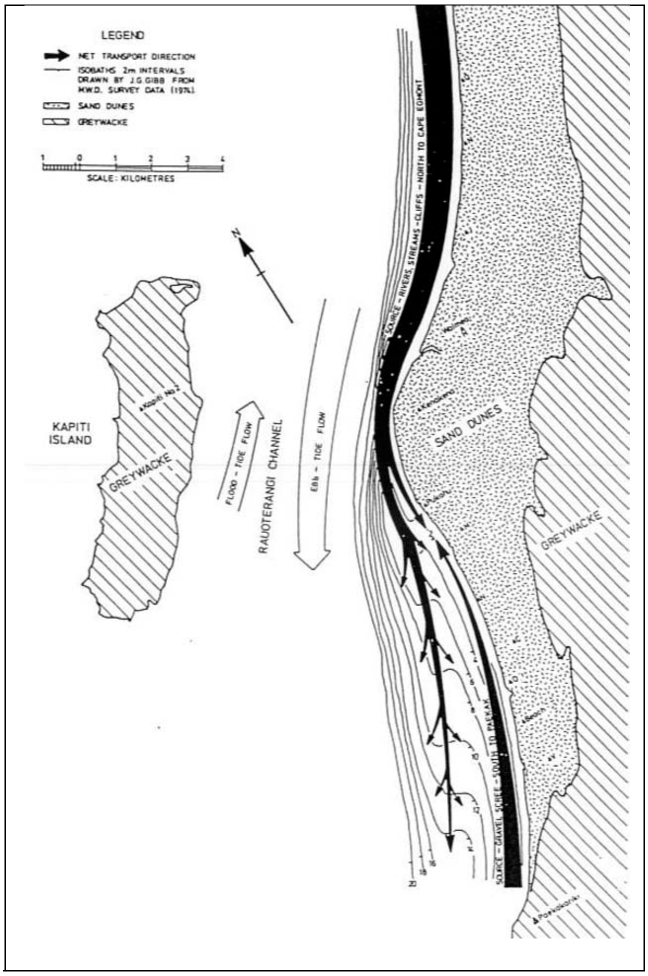


Figure 10. Proposed sediment transport pathways for the southern section of the Kapiti Coast from Te Hapua to Paekakariki (Figure 8 from Gibb, 1978).

The evidence discussed above indicates that the behaviour suggested by Gibb (1978) is broadly correct. However, there is little contribution of sediment from the south. It is more likely that sediment moves onshore during relatively calm low amplitude swell conditions. Hence, the sediment supply for the southern flank of the cuspate foreland is primarily driven by recirculation of sediment ultimately derived from the north. Since the northwards movement of sediment along the coast of the southern flank of the cuspate foreland is predominantly associated with storm waves, it tends to occur episodically at high rates. The replacement of lost sediment will occur at slow rates over longer time periods. Therefore, this section of coast is likely to show a strong decadal cycle of severe erosion followed by prolonged recovery.

Implications for managing 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 just south of Waikawa. The sediment budget is positive, resulting in accretion throughout the Holocene at an average rate of 0.4-0.6 $m.y^{-1}$. Accretion is continuing at present (CSL, 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 (CSL, 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 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 (CSL 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.
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. Gibb (1978) also indicated that accretion had occurred in the central region occupied by Queen Elizabeth II Park. CSL (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. CSL (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. The inlets have had a long history of modifications that vary significantly between inlets, and there are differences between them in terms of the predominant sediment texture and ranges of discharges. Therefore, the inlets should also not be treated as one type of system.

Coastal Systems Ltd methodology

The CSL (2008a & b) 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 includes the key components suggested by various reviews of Coastal Hazard Zonation methodology (Komar *et al*, 1999; Healy and Dean, 2000; Ramsay *et al*., 2012), with no weighting factors for the different components evident in the relationship as expressed in Equation 1 (page 11, CSL, 2008a). An additional combined uncertainty term has been included to give

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

where these were defined by CSL (2008a, 2012) as:

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 cadastral maps and aerial photographs. Strictly, the long-term change should be over a minimum of 60 years to allow for the fluctuations due to climatic oscillations such as IPO and SAM. However, as discussed below, the time interval used was variable, which is presumably why the *LT* term is referred to as *longer*-term relative to the *shorter*-term fluctuations;
3. *ST* = *Shorter*-term historic fluctuation. From the discussion in CSL (2008a) this was to be derived from statistical analysis of historical data, but in practice it was estimated from the residuals of an Ordinary Least Squares (OLS) fit to the longer-term trend. For assessing coastal erosion, this is probably the most important term as arguably sea level rise does not directly cause erosion for sandy coasts, but acts to increase the elevation to which storm processes affect the beach;
4. *SLR* = Shoreline retreat associated with sea-level rise induced by global warming. CSL (2012) renamed the term *RSLR* to represent the shoreline retreat associated with sea level rise. This terminology assumes that future sea level rise can only cause erosion, and therefore *SLR* will be retained for this discussion. Strictly the *SLR* term should be due to the effect of a change in the rate of 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. This term is required if 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. CSL (2008a) defines this 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 CSL (2008a; 2012) to determine each of these terms is considered in more detail below:

LT – Longer-term trend derivation and uncertainty

The longer-term trends were derived from aerial photographs, and pre-digitised shorelines determined by the National Water and Soil Conservation Organisation (NWASCO) predominantly from aerial photographs and un-specified 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 at the time of the survey on the cadastral maps produced an over-estimate of shoreline erosion rates. The two different shoreline indicators may be several to tens of metres apart at any one time, depending on beach state.

According to CSL (2008a) landward reference points were used to define 68 locations, and the distance between the shoreline and reference point measured in GIS (presumably, as it was not stated) from the geo-rectified aerial photographs and NWASCO plotted shorelines.

CSL (2008a) 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. It is not clear if this was determined separately for aerial photos and NWASCO shoreline cadastral data. For each location about 9 measurements were made from aerial photographs, and 1-2 from cadastral map shorelines. These should have different uncertainties, as generally the error would be expected to differ with the scale of the aerial photograph and the technique used.

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

1. Entire record – 1870s to 2007
2. Earlier period – 1870s to early 1950s
3. Later period – 1940s to 2007

These dates are not exact because the survey coverage varies along the coast, so the dates varied with location. The earlier period was assumed to be unaffected by coastal management, but there is clear evidence that the dunes were affected by grazing and burning resulting in extensive vegetation loss and destabilisation (Hesp, 2001; Hilton, 2006). Following the Sand Drift Act (Introduced 1903, enacted 1908) the dunes were planted in *Ammophila* (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 (Grant, 1991). Development of infrastructure of the dunes also began in the early 1900s. However, coastal protection and flood control structures mostly were first installed in the early 1950s.

NZ studies have identified decadal-scale patterns of shoreline fluctuations (de Lange, 2001), and Grant (1981) identified these patterns for the Kapiti Coast. This means that it is necessary to ensure that the influences of decadal-scale fluctuations are removed from long-term trends, and also the probabilities of coastal hazard extremes (de Lange and Gibb, 2000a & b). CSL (2008a) treated “non-linear” trends using break-point analysis without any constraints on the minimum trend duration that would allow discrimination between trends and fluctuations (Figure 3 CSL, 2008a). This approach has a significant effect on the *LT* term required for the analysis. In particular, CSL (2008a) uses this approach to replace long-term (~100 year) trends with trends over only a few decades (*longer*-term). This is demonstrated in Figure 3 of CSL (2008a). In figures 3A and 3C an accretionary trend is transformed into long-term erosion, which is misleading. In Figures 3B and 3D, the magnitude of the trend is altered significantly.

It is claimed by CSL that, apart from the sites in Figure 3 (CSL, 2008a), the later period trend was *qualitatively* similar to the trend over the entire record. No summaries of the longer-term trends were provided. However, summaries of the trends for the earlier and later periods were available in the database. If the later period trend is *quantitatively* similar to the entire trend, then the trends for the two sections should also be similar. Using the data supplied for 47 sites, the ratio of the later period trend to the earlier period trend was calculated, and found to vary from -32 to 815 (Note that a negative sign indicates a switch in trend between periods). This is a very large variation, which is largely due to the effects of 5 sites that have absolute ratios >30. Three sites are at the foreland apex (C13.04 ratio 814, C13.24 ratio 77, C13.44 ratio -32) , one on the southern flank (C3.93 ratio 53), and one on the northern flank (C22.06 ratio 32). One of these sites – C13.44 - was identified in Figure 3B of CSL (2008a).

Six sites (Table 1) appear to have a change in the direction of trend between the earlier and later periods (either from accretion to erosion, or vice versa). Three sites located between the end of the northern Raumati seawall and Tikotu Creek (C10.29, C10.61, and C11.17) and one closer to the Waikanae River (C14.20) show a switch from accretion to erosion. Sites C11.17 and C14.20 are shown in Figures 3A and 3C of CSL (2008a). Two sites located further north show a switch from erosion to accretion (C13.44 and C17.88), and Site 13.44 is shown as Figure 3B of CSL (2008a).

Table 1. Summary of the changes in trends between the earlier and later periods reported by CSL (2008a) for 47 sites assumed to be unaffected by coastal structures along the Kapiti District coastline.

	Accretion to erosion	Erosion to accretion	Consistent accretion	Consistent erosion	Total
Decelerating	3	0	11	2	16
Accelerating	1	2	17	11	31
Total	4	2	28	13	47

The remaining 41 sites retain the same direction of trend, but either they display deceleration (ratio <1) or acceleration (ratio >1). Ignoring the 5 sites with absolute ratios >30, the mean absolute ratio is 2.34 ± 2.43 for the remaining 42 sites. This indicates that the later period overall has increased trends, as reflected by the values in Table 1. However, the earlier period analysis typically combines 1-3 cadastral survey data points with 1-2 aerial photo points, while the later period analysis is entirely based on aerial photo data. Since there is a difference in the shoreline definition between the two types of data that biases the trend, the inferred trends may be erroneous, and it is not clear if the difference in trends between the two periods reflects a real change in rate or an error.

Not evident in Table 1 is that only one site (C32.54 at the Otaki River mouth) has a ratio that lies in the range 0.8-1.2. The implication of these results is that if the LT term was determined in the early 1950s and the same methodology applied to estimate the long-term average shoreline position at the end of the late period, only one site out of 47 would be within ~20% of the actual location. This implication will be examined further in conjunction with the effects of sea level rise below.

Overall, there are significant differences in trends between the two periods analysed, and it is not appropriate to assume that the later period trend is representative of the long-term trend. This is of particular concern because the LT trend is extrapolated into the future by 50 and 100 years, and so small variations in the trend will produce large variations in the CEPD.

CSL (2008a) used a comparison of the earlier period trend with the later period trend to assess the impact of coastal structures, in order to predict shoreline response for scenarios where the structures are removed or fail. It was acknowledged that this approach was problematic, 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 CSL, 2008a). Therefore, it was assumed that in the critical area where structures now exist, the longer-term rate prior to construction was "*stable*". However, this assertion is unsupported by data provided. Instead, Table 1 indicates that few sites were *stable*, and for most the rates of change are different between the two periods.

CSL derives its' longer-term trend from the later period trend, except for those sites with seawalls or a "recent trend change" (Figure 4B CSL, 2008a). Those sites with a recent trend change use a 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 longer-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).

Hence, there is no consistent approach by CSL in determining the long-term trends for the Kapiti Coast. The main approaches for the calculated rates of shoreline movement in CSL (2008a) 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 and probably present along the Kapiti Coast (Grant, 1981; Shepherd in CSL, 2008a).
2. Trends determined by "weighted" OLS for the 1990s to 2007 (non-linear sites) – which is really a short-term trend.
3. "Stable" areas assumed to have no trend due to the presence of seawalls.

Then, if the later period trend is positive (coast is accreting) it is set to zero, unless the weighted OLS trend indicates a recent change to erosion, in which case the recent trend is substituted for the longer-term trend. Hence, a coast that the data and geomorphic evidence shows to be predominantly accreting north of Tikotu Creek is transformed into an erosional coast to assess future risk of erosion.

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 later 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 (SEE), which is the standard deviation of the residuals.
3. Other factors that affect the uncertainty are discussed but then ignored.

CSL (2012) states that "*alongshore smoothing was carried out to derive the 95% confidence band over adjacent transects where similar cross-shore shoreline behaviour was apparent, thus preserving alongshore trends*" (Page 16). This procedure was carried also out for other components in the analysis. It is unclear what was actually done, as the smoothing methodology and derivation of 95% confidence bands is not explained. Further, CSL gives conflicting explanations of the same procedure: CSL (2008a) states "*The maximum (95%) value over several transects with similar characteristics was selected to represent that reach*" (Page 28); and CSL (2012) states "*the* *approach used in the present assessment of applying the upper 95% value for longer-term rates and shorter-term variation derived from several adjacent sectors to all those sectors*" (Page 63).

The different procedures defined all exaggerate the magnitude of the components being considered, as indicated by CSL (2012), which states that the approach used "*may have resulted in an overly large component value being applied to some locations. While general precautionary approaches such as these help to minimize uncertainty and increase the safety margin, they may also result in some hazard distances derived in this report being overly cautious*" (Page 63). The assertion made in this statement that a precautionary approach minimises uncertainty is in direct contradiction with an overly cautious CEPD, and is not substantiated by objective analysis.

The error that should be relevant to the *LT* factor when extrapolating the trend into the future is the uncertainty in the OLS gradient (ie. the uncertainty of *b* in Equation 2 of CSL, 2008a). This indicates how much faster or slower the shoreline could be moving relative to the estimated average rate (ie. the confidence limits for the extrapolation at some specified probability). 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 CSL 2008a). No evidence is presented to support this assertion, but it is clear that for accreting coasts, the methodology produces a rate that bears no resemblance to the measured rate, and appears to be inappropriate.

The report also states that the ±3 m shoreline detection error was found empirically to produce a ±3.7 m error in the actual "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 a one-tailed uncertainty distribution is appropriate and hence the only uncertainty to take into consideration is -3.7 m.

Therefore, setting all accreting coastal sites to zero, and then applying an LT uncertainty of -3.7 m over 50 years transformed the entire Kapiti coastline into an erosional zone (-0.074 $m.y^{-1}$ *cf.* an observed long-term trend of 0.4-0.6 $m.y^{-1}$ for most of the coastline). The results do not reflect the true probability of long-term coastal erosion, or the variation of risk along the coast that is evident from historical shoreline changes.

To summarise, the derivation of the *LT* term for the open coast (CSL, 2008a, 2012), is unreliable for the following reasons:

1. The analysis does not assess a long enough record to determine the long-term trend for the Kapiti Coast. Instead, a *longer*-term trend is based on a maximum of 67 years, and arbitrarily uses shorter intervals if they indicate an erosion trend.
2. The assumption that the later period trend is representative of the longer-term trend is invalid. A comparison of earlier and later period trends indicate that 46 out of 47 sites analysed experienced a different rate of change, and 6 of those also involved a changed direction of change (Table 1). It is not clear if the changed rates of shoreline movement between the earlier and later periods represent a systematic bias in the methodology, a consequence of too short a record to remove 50-60 climatic oscillations, a real change in migration rates, or a combination of all these factors. This indicates that the extrapolation of the derived longer-term trend up to 100 year into the future is very uncertain.
3. By separating the uncertainty from the *LT* term, the analysis incorrectly incorporates components into the *CU* term. In particular, when the accretion rate is set to zero the use of a non-zero uncer-

tainty transforms accreting coasts to an erosional trend. There should be no uncertainty for the application of a constant.

4. The uncertainty for the *LT* term is solely based on the estimated measurement errors for shoreline locations. There is no consideration of the goodness of fit of the OLS trend lines in terms of uncertainties. However, the residual standard deviations are used to estimate the *ST* term as discussed below.
5. Although there is discussion of the use of a 95% confidence band for selection of single values to represent a section of coast (referred to as a *reach*), there is no analysis of the confidence limits of the trends, or the confidence limits of the extrapolated trends.

In conclusion, the *LT* term in CSL (2008a and 2012) does not represent a probabilistic analysis of long-term coastal erosion trends as defined by Ranasinghe *et al* (20120), and hence is not suitable for an appraisal of the risk of coastal erosion.

ST – Shorter-term shoreline fluctuation and uncertainty

The short-term shoreline fluctuation in most coastal erosion hazard assessments accounts for the cut and fill associated with storm events occurring over decadal scales or less. It is generally the most important factor for predicting coastal erosion risk, as it defines the limits of the active beach over decadal time scales. Any structures falling within the shoreline envelope defined by cut and fill cycles can end up within the active beach at some point. If the coast is eroding, the probability of this occurring will increase over time, while the probability will decrease if the coast is accreting. For most of the Kapiti Coast, the probability of being affected by storm cut and fill is likely to decrease in the future due to ongoing accretion.

Analysis of short-term fluctuations 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 recovery 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), such as those that occur between the Otaki River and Te Horo. The recovery phase may also be incomplete if the coastal dune vegetation is disrupted, allowing the beach sediment to migrate inland, as has occurred previously along the Kapiti Coast. Without complete recovery, there will be a net loss from the beach sediment budget, resulting in a longer term erosion trend if there is insufficient longshore sediment transport to replace the loss;
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 have 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. Coco *et al* (in press) observed the impacts of a cluster of storms on the French coast, and concluded that it is not possible to scale up the effects of individual storms to predict the effects of a cluster of storms. The corollary is that it will be difficult to untangle the cut and fill effects of individual storms during a cluster of storms.
4. The impact of storms along a coast is generally not uniform. Depending on the pre-existing geomorphology, some areas can be severely eroded while other areas accrete. Key elements of the geomor-

phology that have been associated with longshore variations in storm erosion are variations in beach state (Amaroli et al., 2013), variations in offshore bar/shoal locations and presence of major rip systems (Komar *et al*, 1991; Stephens et al., 1999, Anthony, 2013), and the continuity and elevation of the foredune system (Houser, 2013).

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.

CSL (2008a) refers to *shorter*-term fluctuations, which appears to indicate a different approach to the analysis of cut and fill cycles. Some beach profile data were available, but were not utilised (footnote page 27 CSL, 2008a). CSL (2008a) provided a range of reasons for rejecting the profile data sets, largely due to difficulties with locating the profiles in relation to the shorelines derived from vegetation cover.

However, after examining the profile data provided by Kapiti Coast District Council, the profiles do appear to be suitable for characterising the short-term fluctuation. The purpose of the *ST* term is to provide an estimate of the variability of the shoreline location about the longer-term trend resulting from cut and fill. Therefore, provided the profiles are sampled sufficiently frequently at specific locations, it should be possible to determine the variation about an average profile. Commonly, the short-term fluctuations are expressed as multiples of the standard deviation (typically 3 to approximate 99% confidence limits assuming a Gaussian distribution) of the profile change at selected elevations. This type of analysis appears to be feasible for the Kapiti coastline.

Instead CSL (2008a) assumed that the shorter-term fluctuations are represented by the residuals between the measured shoreline location and the trend line. Hence, the *ST* term was based on the standard error of the estimates (SEE) for the OLS best-fit line by assuming it is equivalent to the standard deviation of the measured profiles, giving *ST* = ± 3×SEE. However, 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. Although the later period trends determined by CSL (2008a) involve only one type of measurement, there is still a measurement error that is incorporated in the residuals. In particular, the errors in geo-rectification and shoreline position determination appear to be of a similar magnitude to the calculated standard error of estimates (Figure 6A CSL, 2008a; Table 3.1 CSL, 2012);
2. The vegetation lines are not likely to represent the average shoreline position (assumed by the CSL methodology). As noted in CSL (2008a), 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, and there may be a seasonal effect on vegetation extent. Shore profile data may also be biased towards an eroded shoreline, as there is often a tendency to undertake more frequent surveys following a storm, and less when the beach is considered stable or accreting; and
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 as discussed above, so a proportion of the residuals represents fluctuations in the long-term rate.

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 standard deviation of the residuals appears to be the error term considered for the uncertainty of the *LT* factor, and therefore this has been incorporated into the CEPD more than once.

Appendix C of CSL (2008a) compares the estimated *ST* term with the reported cut and fill shoreline changes of Gibb (1978), focussing on his long-term trend data. CSL (2008a) argues convincingly that the large fluctuations in Appendix 1 of Gibb (1978) are due to errors in the shoreline location on early cadastral maps, and therefore the Gibb (1978) short-term values should be ignored. However, the main body of Gibb (1978) bases short-term fluctuations on measured changes during storm events in the 1970s, particularly the 11-13 September 1976 storm, which occurred at the end of a cluster of storms, that produced a maximum of 15 m erosion at the Raumati seawall, and an average of 6 m elsewhere along the coast. This compares to *ST* values from CSL (2008a) ranging from 10 to 36 m, with the lowest values occurring along the southern flank of the foreland, which Gibb reported as having the largest storm cut that he attributed to the influence of seawalls that failed during the storms, and the highest values occurring near the Waikanae River, which experienced much lower storm cut in the 1970s. Overall, the estimated *ST* values of CSL (2008a) appear inconsistent with observed storm cut.

Gibb and Depledge (1980) provide further data on cut and fill for the Paekakariki area for storms that occurred from December 1978 to January 1980, producing maximum storm cut of 7-12 m. Based on the calculated long-term erosion and storm cut, Gibb and Depledge (1980) recommended the immediate removal of 13 NZ Railway houses on the seaward side of the southern end of Ames St, Paekakariki, to be followed by the removal of the next 20 houses further north over 5 years. The first 13 houses were removed from the coast between sites C0.40 and C0.73, while the other properties are still occupied (Appendix A CSL, 2008a). The evacuated properties do not have any seawalls or other coastal protection. Appendix A, and the database provided indicate that there has been a reduction in erosion over the later period analysed by CSL (2008a). However, this includes the erosion from the 1970s that resulted in the house removals. Since the houses were removed, the data indicate stability to slight accretion, contrary to the predictions of Gibb and Depledge (1980).

The uncertainty for the predicted *ST* was derived from the measurement errors related to the OLS determination using an undefined empirical method. This gave an uncertainty of ±2.6 m. For the CEPD summation, only negative values for *ST* and the uncertainty were considered. Again, for the accreting areas of the Kapiti Coast, this approach will exaggerate erosional hazard in the future.

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.

In conclusion, the derivation of the shorter-term trend by CSL (2008a, 2012) uses a method that differs from standard practice, does not appear to be a valid approach, and does not provide a probabilistic assessment of the cut and fill extent. The predicted values appear to be inconsistent with observed storm events.

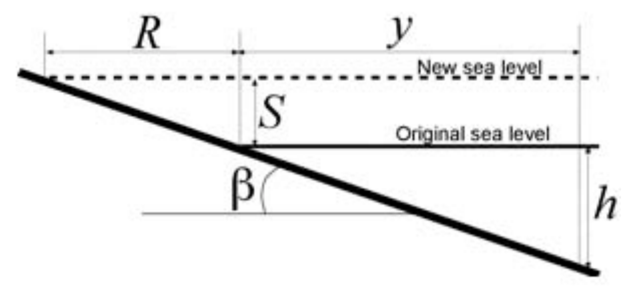
SLR – Impact of sea level rise determination and uncertainty

This factor is included to account for accelerating sea level rise anticipated as a consequence of global warming, and CSL (2012) renamed the term *RSLR* to represent the shoreline retreat associated with sea level rise. Since the available evidence shows no relationship between sea level and shoreline retreat along the Kapiti Coast, this relabelling is inappropriate and reflects an assumption that future sea level rise can only cause erosion. Therefore, the symbol *SLR* will continue to be used in this discussion.

The *LT* factor discussed above already includes the effects of historic relative sea level changes and is extrapolated into the future. Therefore, the *SLR* factor should strictly be based on the additional 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.

For stabilised parts of the Kapiti Coast (with seawalls), it was assumed that sea level rise would not cause erosion while the structures were maintained for up to 50 years (CSL, 2012), while the 100 year predictions assumed all structures were immediately removed. Without a maintained structure, it was assumed that sea level rise would automatically lead to coastal erosion. This assumption is commonly made for the effects of future sea level rise (FitzGerald *et al*, 2008; Ranasinghe and Stive, 2009; Jackson *et al*, 2013). For example Zhang *et al* (2004) suggested that the underlying rate of erosion of sandy coasts is "*two orders of magnitude greater* than the rate of rise of sea level" (italicised in the original). There are some difficulties with this assumption. Firstly it is clear from observations that past sea level rise is not consistently associated with erosion of sandy coasts (FitzGerald *et al*, 2008; Anthony, 2013), and this is currently the case for most of the Kapiti Coast. Secondly the assumption of future coastal erosion is largely based on numerical predictions derived from the *Bruun Rule* (BR) and/or *Equilibrium Beach Profile* (EBP) concepts (SCOR Working Group 89, 1991; Thieler *et al*, 2000; Ranasinghe *et al*, 2012).

Both conceptual models can only predict erosion due to their inherent assumptions about the response of a beach system to rising water levels (Figure 11), which is primarily that there is an upward and landward adjustment of an idealised beach profile (SCOR Working Group 89, 1991; FitzGerald *et al*, 2008). Note that this approach should also predict accretion for falling water levels as occurs on the Kapiti Coast in response to climatic oscillations, such as ENSO and the IPO (Bell and Hannah, 2012), which has not been observed.



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

Figure 11. 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 should be obvious that the rapid influx of sediment onto the coast of the Manawatu that started around 7,000 BP could not have occurred if the assumptions of the BR or EPB models were valid. Although some aspects of the BR and EBP conceptual models have been demonstrated under controlled laboratory conditions, field tests show that these methods have no predictive value. For example, List et al (1997) used the BR and measured relative sea level changes to hindcast the shoreline erosion for Louisiana barrier islands in the USA, and they found no significant correlation. Hence, they concluded that the BR approach has no power for hindcasting or forecasting the effects of sea level rise. Following a series of reviews of the factors driving coastal change for the entire USA and Hawaiian coast, Hapke *et al* (2013) found that geomorphology and human activities were the primary controls on coastal erosion, probably through their effects on the sediment budget. Anthony (2012) found the same for the southern North Sea. Pickett (2004) assessed the use of

EPB models for predicting coastal hazards in the Bay of Plenty, New Zealand. He found no significant correlation between relative sea level rise and EBP predicted shoreline response.

Consequently it is evident that the BR and EBP approaches are unsuitable for predicting shoreline response to sea level rise (SCOR Working Group 89, 1991; Thieler *et al*, 2000; Cooper and Pilkey, 2004; Davidson-Arnott, 2005; FitzGerald *et al*, 2008). CSL (page 32 2008a) agrees that the BR approach is not appropriate and indicates that it shouldn't be used.

Appendix D (CSL, 2008a) discusses models for predicting shoreline response to sea level rise. It confuses the original BR (Bruun, 263; 1983; 1988) with later variations of it, particularly the Weggel (1979) modification, 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 gradient (Weggel, 1979), or the steeper slope of the offshore bar (Dubois, 1977), neither of which are dependent on the closure depth. 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 (Figure 11). The BR method discussed in the report (Equation D1 CSL, 2008a) attempts to approximate this by including the height of the sub-aerial berm or foredune, which is the Weggel (1979) formulation, and a common modification of the BR (Rosati *et al*, 2013).

CSL (2008a) suggests that the Komar *et al* (1999) equation is a better alternative. This relationship was developed to predict the extent of storm cut during a single event, albeit for the largest expected storm over a specified time period. It was developed for the Oregon coast, and Komar *et al* (1999) note that due to tectonic effects parts of the coast are experiencing relative sea level fall, while other areas have a relative sea level rise. They also observed that sea level rise is not a significant factor. Equation (2) in Komar *et al* (1999), which defines the coastal hazard zone, makes it clear that the method is not a function of sea level rise, as a separate term is included for projected sea level rise effects. Equation (3) in Komar *et al* (1999) defines the maximum dune erosion, and can be expressed as (see Figure 12 for definition of parameters):

$$R_{\max} = \frac{(\eta_{\max} - z_{toe}) + \Delta BL}{\tan\beta}$$

This equation predicts the maximum expected dune erosion by assuming that the saturated beach face can be projected inland until it intersects the extreme water level, and all the sediment above that surface is removed by erosion if the extreme water level is above the dune toe elevation (Figure 12). The method also allows for the beach surface to be adjusted for any erosion that occurs during the storm.

The method was tested against available data for dune erosion along the Oregon coast, which seems to have involved dissipative beaches. Further, *Ammophila arenaria* (known as European beach grass in Oregon) was introduced to the Oregon coast in the late 1930s to stabilise drifting sand (Reckendorf *et al*, 1985). It has progressively invaded the coastal dunes, leading to artificially high and continuous foredunes that didn't previously exist (Wiedemann, 1996), similar to *Ammophila* dunes in New Zealand (Hilton, 2006). This suggests that the Komar *et al* (1999) method would be an

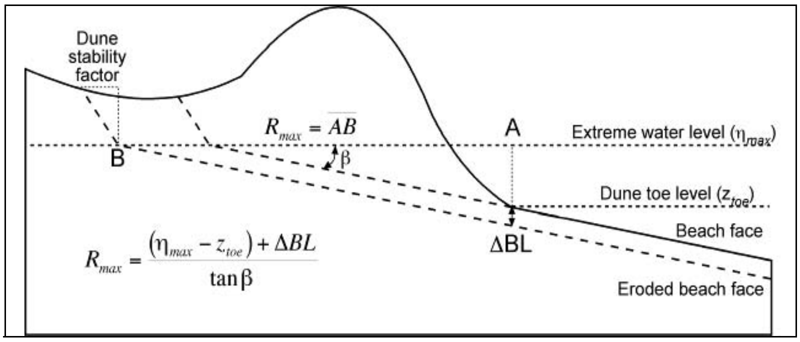


Figure 12. Definition sketch for the foredune erosion model in response to storm events proposed by Komar *et al* (1999)

appropriate approach for assessing the short-term cut (*ST* term) for the *Ammophila* dunes of the Kapiti Coast in conjunction with the extreme wave and water level probability distributions reported by MetOcean Solutions Ltd (2010).

It appears CSL (2008a) modified Equation (3) of Komar *et al* (1991) by replacing the numerator term with sea level rise, indicating that the *SLR* term is equal to the ratio of sea level rise to the slope of the beach (Equation 3 CSL, 2008a). This is the functional form of the BR, particularly the Weggel (1979) modification. Therefore, for all practical purposes tanβ = L/[B+d], so there is no real difference between Equation D1 that CSL (2008a) correctly argues should not be used, and Equation 3 that CSL (2008a) did use.

The method used by CSL (2008a) 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 where repeated profile measurements were available. It seems that the available profile slopes were averaged, but it is not explained how it was done or what the variation about the averages were. The calculated slopes were rounded down in order to increase the predicted shoreline retreat. The profile sites did not coincide with the coastal hazard calculation sites, and so slopes were interpolated. No errors were defined for the interpolated slopes.

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 the order of 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 global temperatures. So far this has not been the case (Gregory *et al*., 2012), and more than 40 years of sea level projections have not successfully predicted the actual global sea level response (Gehrels 2010; de Lange and Carter, 2013, Houston, 2013). Most studies have found that the global rate of sea level rise determined by long-term tide gauge records has been decelerating for at least the last 50 years (de Lange and Carter, 2013), and this is also evident in the shorter, more recent satellite record (Chen *et al*., in press).

At a regional scale, the projections for the Tasman Sea significantly overpredict the observed sea level rise (Boretti, 2012). Finally for the local Kapiti Coast, the measured relative sea level rise of 2.03 $mm.y^{-1}$, which includes the effects of tectonic subsidence (Bell and Hannah, 2012), is lower than the 3.7 $mm.y^{-1}$ sea level projections assume the rate has accelerated to by 2013 (IPCC WGI Fifth Assessment Report – Chapter 13). Assuming that the difference between the observed and assumed rates for the Kapiti Coast remains constant for the next 50 years, it would equate to a difference of 0.8 to 6.0 m for predicted *SLR* term. If the observed sea level rise accelerates at a lower rate than assumed for the projections, the difference will be larger, and if the observed deceleration in the rate of sea level rise continues, the difference will increase still further.

Figure 13 compares the average shoreline response (ignoring the *ST* and *DS* terms) for the period 1950-2007 assuming the observed rate of relative sea level rise for Wellington of 2.03 $mm.y^{-1}$ (Bell and Hannah, 2012). This value is higher than the 1.7 mm.y[-1] reported by CSL (2008a) due to subsidence of the Wellington region associated with slow slip earthquakes (Beavan and Litchfield, 2012). Since it is the observed rate at Wellington, it may be a little too high as the effect of the observed subsidence is smaller for Kapiti than Wellington (Beavan and Litchfield, 2012).

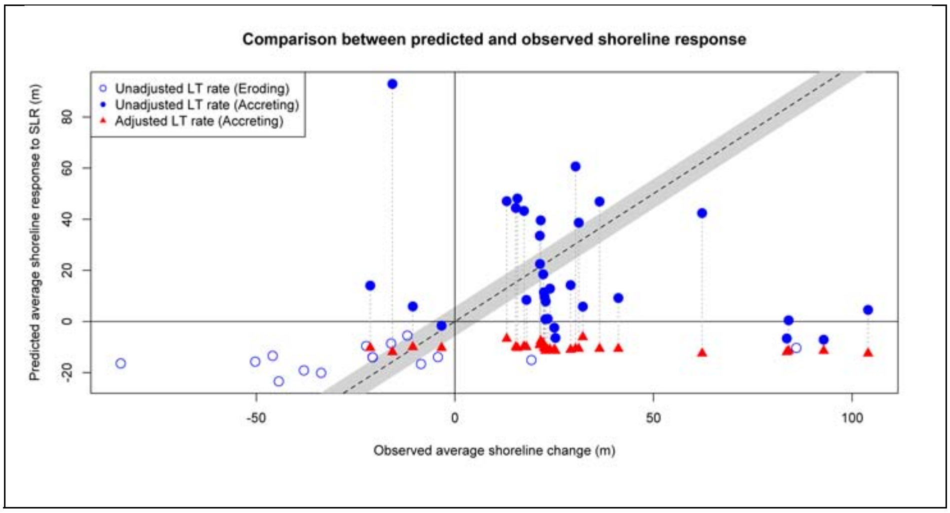


Figure 13. Comparison between the predicted and observed average shoreline change between 1950 and 2007 using the early period *LT* erosion/accretion and the *SLR* erosion determined by the BR method for a rate of sea level rise since 1950 of 2.03 mm.y[-1] (Bell and Hannah, 2012). The adjustment for accreting coasts used by CSL (2008a) was also applied to locations accreting during the early period (red triangles connected to unadjusted predictions by vertical dotted lines). The shaded grey zone indicates agreement between predicted and observed shoreline response allowing for a *CU* term of ±6 m.

The effect of setting the *LT* term to zero for accreting coasts and applying a non-zero uncertainty was also assessed for locations that were accreting during the early period considered by CSL (2008a). Vertical dotted lines connect the predicted shoreline locations without (red triangles) and with (solid circles) an accreting shoreline. The sloping dashed line indicates perfect agreement between predicted and observed coastal erosion, with the grey shading indicating the *CU* uncertainty adopted by CSL (2008a) of ±6 m. It is evident that, using historical data for sea level rise, there is poor agreement between predictions and observations. Further, the adoption of a zero trend for accreting coasts does not improve the agreement, as there were 3 sites within the grey zone before adjustment, and 2 different sites after adjustment. Overall, the methodology of CSL (2008a) provided hindcast predictions within the specified uncertainty for 4-6% of the cases, which does not provide confidence in the predictions for the future.

The hindcast analysis used a known sea level rise, but this is not known for the predictions of the future. Therefore, for an assessment of risk it is of concern that there are no probabilities 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. CSL (2008a) is based on a value of 0.6 m.Century[-1], which is three times the observed rate of relative sea level rise for Wellington since 1944, while CSL (2012) used 0.6 m.Century[-1] for the 50-year projection (0.3 m total) and 0.9 m.Century[-1] for the 100 year projection. The assumed sea level rise was described as conservative (page 34 CSL, 2008a).

There is no indication of the probability of occurrence for the assumed sea level rise, which is required for risk assessment. Considering the IPCC AR4 projections (IPCC, 2007), used to develop the Ministry for the Environment guidelines for New Zealand, and the more recent IPCC AR5 projections[1], the worst case, and hence least likely, scenarios are suggesting maximum sea level rises of 0.6-0.8 m.Century[-1] with mid-point sea level rises of 0.4-0.6 m.Century[-1]. Hence, the sea level rises used by CSL (2012) are higher than those summarised by the IPCC

[1] The IPCC AR5 projections are currently only available in draft form and may be changed to align with the published Summary for Policy Makers before being published in 2014.

(2007, in press), and involve rates of sea level rise that have previously only occurred for short durations during meltwater pulses following the Last Glacial Maximum (Standford *et al*., 2011). Therefore, the probability of the assumed sea levels occurring is likely to be extremely low.

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. It is unclear why the slope measurement was converted to an angle for this determination. The slope error was originally ±0.001 grad, and, since this calculation effectively takes the reciprocal of the slope, the error analysis should have been based on percentage error. The uncertainty should also consider the variability of the nearshore slope, particularly since the method is based on the most variable part of nearshore geomorphology. The *SLR* uncertainty should consider the uncertainty of the sea level projections as well as the slope measurements.

In conclusion, the *SLR* term was determined by an inappropriate methodology that incorrectly determines the response to sea level as demonstrated by hindcasting 57 years of shoreline change for the Kapiti Coast (Figure 13). No analysis of the probability distributions of the key parameters used was undertaken, and therefore, the results cannot be used in a risk assessment.

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 quite common, and assumes that the material falling from the top of the slope accumulates at the toe until a stable slope is achieved. CSL (2008a) assumes that half of the stable slope occurs landward of the dune toe at the end of the storm, and the other half occurs seaward. Hence, the *DS* term is half of that often applied by assuming the stable slope is located entirely landward of the storm dune toe. The result depends on the assumed stable slope angle and the height of the scarp. CSL (2008a) assumed a stable angle of 34° for the Kapiti Coast dunes, while noting that some stable dune scarps around Paekakariki have angles of 41°. In contrast, Gibb and Depledge (1980) assumed a stable angle of 40° based on measurements after storm scarping of the dunes at Paekakariki that ranged from 35° for loose dry sand to 45° for vegetated damp dunes. Therefore, the *DS* term is likely to overestimate the retreat required to produce a stable slope.

CSL (2008a) assumed that the scarp height resulting from future erosion equated to the maximum dune height along sections of similar coast including each location 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 termination is coincident with the maximum dune height. Overall, the approach used will over-estimate *DS*, as noted in the report (page 36 CSL, 2008a).

The uncertainty is based on the root mean square (RMS) 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, which is likely to underestimate the steepness of the dune scarp as observed by Gibb and Depledge (1980).

In conclusion, the *DS* term should not have been included in the CEPD assessment. Further the methodology used over-estimates the *DS* term, although this is offset by assuming that only half the *DS* term occurs landward of the dune toe at the end of the storm.

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*. However, there do not appear to be any such factors actually included in the *CU* term.
3. The methodology repeatedly selects values that maximise the possible erosion as a conservative or precautionary approach. There is no analysis of the extent to which this increases the final CEPD, or what the CEPD would be if alternatives that minimise coastal erosion were used.

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 (Page 38 CSL 2008a), but it shouldn't be included and it does not appear to have been. It was also stated that the 5 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, and Equation 5 indicates *CU* is a function of the other terms.

The calculated *CU* factor was ±5.3 m, which was rounded up to ±6 m for the 50 year CEPD (CSL, 2008a). It is clear from Figure 13 that this underestimates the errors in the predicted shoreline changes. CSL (2012) recalculated the *CU* factor for the 100 year CP, and obtained ±9.5 m, which was rounded up to ±10 m suggesting an increased confidence in the results for the second half of the century.

CSL (2012) also lists a number of contributions to uncertainty that were considered unnecessary to be included because the conservative and precautionary methodology already over-estimated the erosion, and that this compensated for the uncertainty of the projections of future climate. This is an unusual approach to quantifying uncertainty, and seems to advocate a particular planning position on acceptable risk rather being an objective approach to risk assessment.

It is evident that at each step of the determination of the CEPD, the analysis maximises the estimated future shoreline erosion, and the effect it had on the resulting CEPD has not been quantified. Of particular concern is that this approach ignores any mitigating factors, except for the presence of some seawalls. Overall, it has the effect of exaggerating the future hazard and almost certainly has identified areas as being hazardous that are unlikely to experience any coastal erosion. Therefore, it represents an unrealistic assessment of the potential risk associated with coastal erosion.

Removal of structures

CSL (2008a, 2012) also has predicted the CEPD for locations currently protected by seawalls based on three scenarios:

1. The seawalls maintain their current level of protection for the duration of the prediction period (*Seawalls hold*);
2. The seawalls occasionally fail, but are quickly repaired or replaced (*Seawalls repaired*); and
3. The seawalls fail and are removed at some stage during the prediction period (*Seawalls removed*). This scenario was omitted from the 2012 update (CSL, 2012).

Somewhat confusingly, the three scenarios were also applied to regions with no seawalls, but using a different methodology. Only the coast south of Marine Parade, Paraparaumu Beach (site C11.17) appears to have sites where the three scenarios have some relevance. CSL (2008a) also distinguishes between *official* and *private* seawalls, where official seawalls were built and/or maintained by the Kapiti Coast District Council and protect multiple properties and public land. Private seawalls are built and maintained by private individuals, and it is assumed that they only provide partial protection. It is not clear if the distinction resulted in a different methodological approach.

CSL (2008a) does not clearly explain the methodology for the different scenarios, stating that the methodology was defined in the database. The approach appears to have been:

1. *Seawalls hold* methodology set all the terms to zero, so the CEPD is zero;
2. *Seawalls repair* methodology assumes that there is some coastal erosion before the seawall is repaired, and this erosion consists of *ST*, *DS*, and *CU* terms. The *ST* term was interpolated from adjacent non-seawalled sites and seems to be 15 m for most sites. The *DS* term was calculated from the local maximum dune height, and the *CU* term was increased to ±9 m to account for scour in front of the damaged seawall. It is assumed that the maximum possible erosion occurs regardless of the extent of damage, or the duration of the repair.
3. *Seawalls removed* methodology includes all the same terms as used for an unprotected shoreline. The only difference is the calculation of the longer-term rate, which represented the sum of an estimated rate of shoreline change if the seawall had not been constructed (LT_{50}) and a catch-up allowance for the amount of erosion that may have occurred over the past 50 years if the seawall was not present (LT_{cu}). For a 50 year prediction period, the two components are equal ($LT_{50} = LT_{cu}$) so (CSL, 2008a) effectively replaced LT with $2\times LT_{50}$. This approach implies that for a 100 year prediction the LT would be $3\times LT_{50}$ (100 years of the long term trend plus the 50 years of catch-up). However, the values given in Appendix D (CSL, 2012) correspond to $4\times LT_{50}$. The longer-term rate used to calculate LT_{50} was based on the calculated earlier period trends. The calculated trends for adjacent sections of coast were smoothed and the 95% maximum erosion rate estimated (Accreting trends were set to zero). The erosion rate was then rounded to the nearest 0.05 m.y^{-1} to allow for the less reliable cadastral-based data for the earlier period rate. CSL (2012) discusses an alternative approach based on the behaviour of the unprotected coast between Paekakariki and Raumati South, and concluded that the CSL (2008a) methodology was appropriate.

Overall, the methodology used is likely to over-estimate the shoreline erosion, particularly in the case of the *seawalls removed* scenario. This arises due to over-estimation of the erosion rates, and also because of the assumption that the erosion occurs for the full duration of the prediction period (no consideration of when the seawalls are removed), or to the maximum possible extent during a seawall repair with no mitigation measures to minimise erosion, to repair the effects of erosion.

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 (Hart, 2009a & b). 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.

CSL (2008b) argues that for the Kapiti Coast, the hazards associated with tidal inlets are significantly different to those experienced on the intervening open coasts. This is reasonable in that inlet migration only occurs at inlets, and requires that the future behaviour of the inlet be reliably predicted. The open coast CEPD equation was modified by replacing the short-term fluctuation with an inlet migration factor (*IM*) to account for inlet migration (CSL, 2008b, 2012). Note that the subtraction operation in the equations defining the *IEPD* (Equation 2 CSL, 2008b; Equation 6 CSL, 2012) is incorrect, as the terms are all added together to define the landward movement of the shoreline. The *IEPD* merely replaces the short-term fluctuations due to wave process (*ST*) with the short-term fluctuations associated with channel migration. It does not take into account any of the other hazards, such as flood inundation, that may be associated with inlets.

To determine the inlet migration CSL (2008b) selected points that represented the maximum landward excursions evident in aerial photographs since 1939 based on the location of vegetation regardless of longshore position. This doesn't really correspond with accepted interpretations of inlet migration that relate to the longshore stability of the main channel (viz. Hayes, 1980; Komar, 1996; Hart, 2009b). It is difficult to envisage how the CSL (2008b) approach will provide suitable data for a probabilistic analysis of coastal erosion risk.

Further, by using vegetation to indicate shorelines, there is likely to be a significant lag between the migration of the shoreline and establishment of vegetation, particularly if grazing and other anthropic factors are present. Earlier cadastral surveys, which were based on the position of the high tide mark, were only used to estimate the location of the main inlet channel(s).

The maximum landward excursions from the entire set of inlet shorelines measured were then combined to produce a *composite shoreline*, which represents the maximum landward extent of the envelope of all inlet shoreline positions. Note that at no time during the period of analysis did the inlet shoreline simultaneously occupy all positions along the composite shoreline. The composite shoreline is then transformed into the *inlet migration curve* (*IMC*) by fitting a curve that was "*consistent with the general shape*" (page 15 CSL, 2008b) of the local maximum landward inflexion points along the composite shoreline. Finally the *LT*, *SLR* and *DS* terms from the nearest open coast site were used to calculate an offset that was combined with the inlet *CU* term (*LT*+*SLR*+*DS*+*CU*) to shift the inlet migration curve inland to become the *IEPD*.

The uncertainty term *CU* for the inlets used by CSL (2008b) should differ from the open coast *CU* term (CSL, 2008a) due to the substitution of the *ST* term with the *IMC*. CSL (2008b) calculated the *IMC* uncertainty solely from the measurement error of the digitised inlet shorelines, and determined a total *CU* over 50 years of ±5.9 m (*cf.* ±5.4 m for the open coast), which was then rounded up to ±6 m, matching the open coast *CU* value adopted by CSL (2008a). Similarly CSL (2012), derived an inlet *CU* term of ±10 m over 100 years that matched the open coast value. There was no quantification of the uncertainties involved in the conversion from measured shorelines to the inlet migration curve. In particular, the fitting a curve to approximate the general shape introduces additional errors not account for by the *CU* term. Therefore, the uncertainty is likely to be larger than indicated by the *CU* term.

CSL (2008b) distinguished between *unmanaged*, *transitional*, and *managed* in analysis periods (summarised in Table 2 below). The distinction between unmanaged and managed inlets was on the basis of the inferred effectiveness of any inlet management structures and/or procedures such as the deliberate breaching of any berm blocking the inlet as permitted for many of the inlets by the Greater Wellington Regional Council. Transitional inlets represented time periods where the effectiveness of management was uncertain. Data for transitional periods were excluded from the derivation of inlet migration curves. CSL (2008b) further distinguished between the northern and southern sides of inlets primarily on the basis of the interpreted behaviour of the open coast, the presence or absence of open coast structures, or the potential influence of structures or inlets updrift of the inlet.

It is evident that the application of the methodology varied between inlets by considering different time periods, and the interpretation of the influence of structures, management regimes such as barrier breaching, and the influence of coastal processes. The methodology for determining the inlet migration curve was modified at Mangaone Stream to account for an assumed change in beach morphology. Finally the methodology was also altered for Whareroa and Wainui Streams to incorporate the effect of open coast seawalls not specifically part of the inlet system.

Table 2. Summary of the analysis periods used by CSL (2008b) for the inlets along the Kapiti Coast.

Inlet	Unmanaged period	Transitional period	Managed period
Waiorongomai Stream - North	1942 - 1965	1965 - 1972	1972 - 2007
Waiorongomai Stream - South	1942 - 2007		
Waitohu Stream	1942 - 1966	1966 - 1973	1973 - 2007
Otaki River	1939 - 1946	1946 - 1957	1957 - 2007
Mangaone Stream	1948 - 2007		
Hadfield Stream	1948 - 2007		
Waimeha Stream	1942 - 1966	1966 - 1973	1973 - 2007
Waikanae River	1942 - 1966	1966 - 1980	1980 - 2007
Tikotu Creek	1942 - 1965	1965 - 1972	1972 - 2007
Wharemauku Stream - North			1952 - 2007
Wharemauku Stream - South	1942 - 1966	1966 - 1973	1973 - 2007
Whareroa Stream	1942 - 2007		
Wainui Stream	1942 - 2007		
Waikakariki Stream	1942 - 1956	1956 - 1979	1979 - 2007

An interesting aspect evident from the discussions of the historical development of the inlets in CSL (2008b) is the progressive southward appearance of a pulse of sediment affecting the inlet morphology. This is reported for the northern-most inlets as starting in the 1940s, affecting the Waikanae River in the 1950s and 1960s and finishing at the southern-most inlets in the 1970s. Further, it is suggested that it represents a 50-60 year quasi-cyclic process, which is consistent with the findings of Grant (1981) and corresponds to the IPO oscillation modulation of precipitation and wind climate, with a lagged influence along the coastline associated with the rate of longshore sediment transport. The data presented also suggest that another pulse of sediment has been affecting the northern-most inlets for at least the last decade.

The methodology has several problems:

1. The aggregating of multiple inlet shorelines into a shoreline envelope to define the composite shoreline ignores the behaviour of the inlet over time, which means that there are no probabilities associated with shoreline locations. This makes it impossible to assess the risk of erosion. It also obscures any systematic patterns of behaviour that could be used to predict the future pattern of inlet migration.
2. The composite shorelines, and more importantly the IMCs derived from them, do not appear to consider the geomorphology consistently. For example, CSL (2008b) adjusted the IMC for the southern side of the Mangaone after assuming that the 1948 shoreline was in response to a lowered beach berm height. However, there is no allowance for the dunes formed since 1948, which would restrict shoreline erosion.
3. The analysis depends on the determination of what constitutes a managed or unmanaged inlet. The historical summaries presented (CSL, 2008b, 2012) indicate that all of the inlets have been modified in various ways and extents throughout the entire analysis period, particularly the period of aerial photography. It seems that the distinction is based mostly on an arbitrary assessment of the type of structures built within the inlet, presumably to fit with the seawall scenarios on the open coast. There is no analysis of the impacts the structures have on the probability of inlet erosion, apart from recognition that they may restrict inlet migration.
4. Although the data show that most of the inlets occur on accreting coasts, it is assumed that the inlet migration curve can shift landwards in the future.
5. The overall analysis appears to be sensitive to the availability and quality of the data, and the choices made by the analyst.

The Waimeha Inlet (Figure 14) demonstrates the last issue. CSL (2013) undertook a reassessment of the northern side of Waimeha Inlet. This reassessment included aerial photographs taken in 2010 and 2013, higher quality aerial photographs for 1973 and 1988, and additional historical data on inlet modifications. The reanalysis considered two different time periods for the transition between managed and unmanaged inlet conditions – the original from CSL (2008a, 2012) given in Table 2, and an alternative transition period of 1980-1988. This corresponds to 3 different

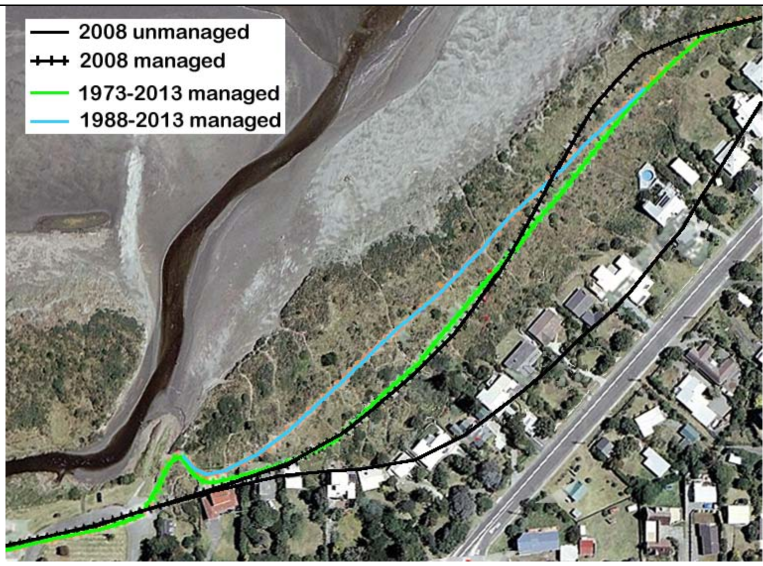


Figure 14. Comparison between the original CSL (2008a) predictions (unchanged in CSL, 2012), and the revised CSL (2013) predictions for 50-year CEPD lines of the northern side of Waimeha inlet. See text for discussion of the revisions.

predictions of the 50 year managed shoreline based on different time periods: 1973-2007; 1973-2013; and 1988-2013.

Figure 14 shows the changes between the managed CEPD lines derived from the three different time periods considered to be affected by inlet management (2008 managed CEPD was based on the 1973-2007 time period). It is evident that the combination of shorelines used to create the inlet migration curve significantly affects the outcome. In particular, the exclusion of the 1980 and 1988 shorelines appears to be the sole factor causing the difference between 50-year managed shorelines based on the 1973-2013 and 1988-2013 periods (Figures 14 & 15)

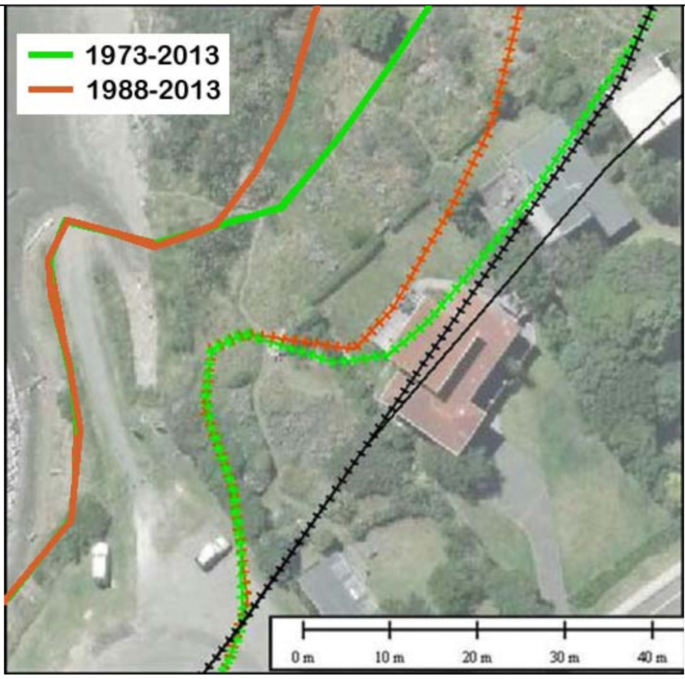


Figure 15. Close-up of northern side of Waimeha inlet showing the CEPD lines (black solid line and hatched lines), with the *IMC*s and managed shorelines from Figure 2 of CSL superimposed (red/green solid and hatched lines respectively).

It is also evident that the interpretation of the influence of structures, as the groyne and stormwater outlet located to the right of the sharp bend in the Waimeha Stream as it exits onto the beach was ignored in the earlier assessments due to the poor quality of the aerial photographs and not being noticed during the site visit (CSL, 2013). The inclusion of these structures produces the bulge on the seaward side of the property at 21 Field Way.

Comparing Figures 1 and 2 from CSL (2013) also suggests that there is an issue with the implementation of the methodology using GIS (Figure 15). The methodology states that the CEPD is a landward translation of the *IMC* by a distance determined by the sum of the other terms, but doesn't explicitly state the orientation of this displacement. For the example in Figure 15, the CEPD is offset in the longshore direction. It would also be legitimate to question why the CEPD doesn't coincide with *IMC* for the region that is protected by a hard structure. Finally, Figures 14 and 15 also demonstrate how the choice of data and how it is included has different impacts on properties in the affected region.

Summary of methodological concerns

The preceding sections outlined concerns with various aspects of the methodology used by CSL (2008a, 2008b, 2012 and 2013), ranging from serious to minor. It should be self-evident that the CEPD lines produced are the consequence of a series of assumptions made and the specific methodology used to derive them. There are three main aspects that invalidate the CEPD lines for the purpose of providing an assessment of the coastal erosion hazard for the Kapiti Coast:

1. At almost every step of the analysis, a procedure was followed that maximised the predicted erosion, which was justified as being a precautionary or conservative approach. The only exception was the choice to distribute the effect of the dune stability factor on either side of the predicted storm erosion extent. However, the dune stability factor was also inflated by the choice of dune scarp height, stable

angle and the inclusion of an estimated scour factor for the seawall repair scenario (dropped for CSL 2012). Since the shorelines were derived from vegetation lines at the top of the dune scarps, the dune stability factor should have been omitted. Consequently, the CEPD represents an extremely unlikely worst-case scenario. Hence, while it may be reasonable to assume that areas landward of the CEPD will not be affected by coastal erosion, it is unreasonable to assume that all areas seaward of the CEPD will be affected. Also, this procedure means that the analyst is deciding what is an acceptable level of risk at each stage of the procedure, rather than those responsible for coastal management.

2. The methods used are inappropriate for the purpose. Aspects of particular concern are:
   a. The *longer*-term trend (*LT*) is based on too short a time period to separate the long-term trend from fluctuations associated with the IPO. Further, the assertion that the later period trend is qualitatively consistent with the overall trend is demonstrably incorrect.
   b. By separating the uncertainty from the *LT* term, the analysis incorrectly transforms accreting coasts to an erosional trend. The use of very short sequences of data to represent the long-term trend is not justified. There is no basis for expecting a sudden reversal of the observed long-term accretion trends.
   c. The derivation of the shorter-term trend from the standard deviation of the residuals from the OLS fit for the longer-term trend differs from standard practice. It does not appear to be a valid approach, and the predicted values appear to be inconsistent with observed storm events.
   d. Available shore profile data would provide a better estimate of the likely cut and fill response for the Kapiti Coast.
   e. The *SLR* term is derived using a common variant of the Bruun Rule, despite it being recognised that the Bruun Rule should not be applied.
   f. The *DS* term should not have been included because the shorelines used in the analysis were based on vegetation lines, and therefore already incorporate the effects of slope instability.
   g. Using the methodology to hindcast the shoreline response over 57 years indicates that the method is a very poor predictor of the observed response. A simpler and more effective method is to extrapolate the long-term trend covering all available data.
   h. The inlet IEPD is based on an assumed landward inlet migration, and not the longshore migration of the inlet that would normally be used to assess inlet stability.
   i. The landward inlet migration is derived from an envelope of shoreline positions. The methodology used is very sensitive to the selection of which shorelines are included, and the assessment of the effects of any structures present. Overall the method for inlets does not seem robust or reliable.
   j. The uncertainty terms are largely based on measurement errors and do not consider errors introduced by the methodology followed. The terms used are not strictly independent, there are unexplained empirical derivations, and values are arbitrarily inflated to account for unspecified uncertainties. Only single-sided *CU* terms are applied to the final CEPD and IEPD lines.
   k. The analysis does not include a probabilistic analysis of the components of the CEPD or IEPD, and hence cannot form the basis of a coastal erosion risk assessment.

3. Apart from the distinction between the open coast and inlets, the methodology is assumed to apply to the entire coast. There is good evidence to show that the behaviour of mixed-sediment beaches is significantly different to that assumed for sandy beaches. This affects the coast between the Otaki River and Te Horo Beach, and the southern area of Paekakariki to a lesser extent. There is a growing body of evidence that dunes with established native vegetation respond differently to storm events than those stabilised by introduced *Ammophila*. Further, *Ammophila* affects the inland loss of sediment from the coast. As community initiatives are replacing *Ammophila* with native dune species along the Kapiti Coast, the response to coastal forcing is changing and should be accounted for with more suitable methods. Overall, it is evident that a single methodology for the entire open coast is not appropriate.
4. A risk assessment of coastal erosion should include a probabilistic analysis of the drivers and responses for the coast. In terms of drivers for coastal erosion, the analysis adopts values for sea level rise that are suggested for consideration by the Ministry for Environment 2008 guidelines, but does not consider their applicability or probability of occurrence. The analysis assumes that the future climate will adversely affect sediment supply to the Kapiti Coast, but does not quantify the probability of this occurring. It should be noted that the NIWA climate projections (http://www.niwa.co.nz/our-science/climate) do not show any significant change in the coastal drivers other than sea level before 2050, and there is low to moderate confidence in some change by 2090, but the regional effects are very uncertain. Having assessed the probability of changes to the processes driving coastal erosion, the analysis should also have quantified the risk of coastal erosion, allowing for existing mitigating factors. This would provide the necessary data to assess the risk to coastal areas, and also permit a cost-benefit analysis for any proposed management responses.

CSL (2012) recognised that some of the CEPD and IEPD lines were "*overly cautious*" (Page 63). However, it is evident that, due to the methodology followed, all the CEPD and IEPD lines represent an extremely unlikely worst-case scenario. Further, the available data for the evolution of the Kapiti Coast indicate that the shoreline migration is largely determined by the sediment budget, and this budget has been influenced by decadal scale variations in storm activity and not by changing sea level. Climate projections for the next century do not indicate any major changes in storm activity for the Kapiti Coast. Therefore, it is unlikely that significant changes in sediment budget, and thus shoreline migration, will occur in the next century. Hence, the observed changes over the past century, allowing for the effects of structures and management practices, will be a good indicator of coastal erosion hazard (as demonstrated by comparing earlier and later period shoreline trends).

Based on this reasoning, areas experiencing historic shoreline accretion are unlikely to experience an erosion trend in the future, and hence are low risk. In contrast, areas experiencing historic erosion are not likely to experience significant accretion trends in the future, which would make them high risk. However, as noted in CSL (2008a, 2008b, 2012) those areas where historic erosion has affected properties have been modified to mitigate the risk, either by the construction of structures, or the removal of affected infrastructure. Therefore, unless it is policy to remove structures, the future risk is low. Examination of the CEPD and IEPD lines indicate that the majority of properties seaward of the lines occur in areas of accretion, or have protective structures. Hence, it can be concluded that the majority of properties are low risk.

In order to better quantify the actual level of risk, a probabilistic approach should be applied, as discussed below.

Alternative approach

From the available evidence of the Holocene evolution of the cuspate foreland summarised above, and historical shoreline changes for the Kapiti Coast (Gibb, 1978; CSL, 2008a & b), the primary driver of shoreline accretion or erosion is the available sediment (*net sediment budget*). The sediment budget is affected by variations in sediment supply, primarily in response to climatic fluctuations in rainfall and windiness, and to a lesser degree by anthropic factors such as land-use changes and sediment extraction (*viz.* Grant, 1981, 1991). Local sea level variations due to eustatic sea level changes do not have any identifiable impact on shoreline location. Abrupt, large relative sea level changes due to local earthquakes appear to have relatively minor effects on open coast shoreline position, but may affect inlets and can alter the accommodation space for sediment deposition. Local earthquakes can be associated with large increases in sediment supply (Goff *et al*, 2008) and local tsunami, which have probably caused significant changes to the coastal geomorphology of the Kapiti District in the past (Goff *et al.*, 2007).

Given the importance of the coastal sediment budget, an alternative approach would be to first determine the sediment budgets for sections of the Kapiti Coast corresponding to the major geomorphological units. Walton Jr *et al* (2012) review sediment budget methodologies and propose a simplified approach for inlets that can also be utilised for the open coast, although the purpose of their analysis is to identify what can be achieved with a sediment budget.

Table 3 below summarises the data available for assessing the overall sediment budget for the Kapiti Coast. The main sources and sinks of sediments were discussed above in relation to the Holocene evolution of the coastline. Gibb (1978) estimated the volume of sediment required to renourish the Paekakariki and Raumati coast in response to the observed erosion. His estimates correspond to 64 t/m/m of sediment (mass of sediment per metre of beach width per metre of shoreline advance or retreat). This is an under-estimate as it didn't consider the sand volume in the dunes, but gives a reasonable indication of the magnitude. However, taking this value over the entire Kapiti Coast, the observed rate of accretion represents 1.2 kt.y^{-1}. Hence, it is likely that the observed shoreline changes involve mass transport at least an order of magnitude smaller than the potential sediment input to the system.

Table 3. Possible components of a sediment budget for the Kapiti Coast.

<table>
<tr><th colspan="2">Sediment inputs</th><th>Sediment outputs</th></tr>
<tr><td colspan="2">Longshore drift — 80-240 kt.y^{-1}</td><td></td></tr>
<tr><td>Regional
Rivers — 170 kt.y^{-1}
Coastal erosion — unknown
Inner shelf — unknown</td><td>Local
Rivers — 28 kt.y^{-1}
Coastal erosion — unknown
Inner shelf — 0 kt.y^{-1}</td><td>Local
Shoreline advance — 1.2 kt.y^{-1}
Inland — unknown
Offshore — unknown</td></tr>
</table>

Although there are components of the sediment budget missing from Table 3 because they could not be estimated from the literature assessed for this report, they are either relatively easy to assess, such as from comparisons of hydrographic charts for the offshore sediment outputs, or likely to be smaller than the uncertainties in the river sediment inputs. The available data do indicate that a substantial change in the sediment budget would be required to transform the entire Kapiti Coast to an erosional coast.

The sediment budget can be refined by considering smaller sections of the Kapiti Coast, particularly to assess the effects of the 12 inlets along the coast. This would clearly identify areas that have sufficient input of sediment to offset any potential future tendency towards long-term erosion. It would also be useful to assess the effects of sediment pulses moving along the coast. It is expected that such an analysis would replicate the existing pattern of erosion and accretion reported by CSL (2008a), rather than the predicted patterns of coastal erosion implied by the CEPD and IEPD lines.

For areas that are accreting and have a significant surplus of sediment, the CEPD should be predominantly a function of the short-term fluctuations associated with storm events. The extent of erosion can be determined from profile measurements, which is preferred because it would permit a probabilistic analysis, or by the application of analytical models such as Komar *et al* (1991) or Larson *et al* (2004), or numerical models such as XBeach (Roelvink *et al*., 2009). The long-term trends would only need to be considered if there is an intention to continue development seaward of existing property boundaries.

For areas that are eroding, or are identified as likely to experience a sediment deficit in the future, there should be a probabilistic analysis of the CEPD using a process-based model. Ranasinghe *et al* (2012) provide an example of such an approach for Narrabeen Beach, Sydney, Australia, that would be applicable to the Kapiti Coast. The key steps of such an analysis for the Kapiti Coast, assuming the shoreline corresponds to the dune toe, are:

1. Use a Monte Carlo simulation to generate a time series of storms for the future interval of interest using observation based joint probability distributions of the storm characteristics. MetOcean Solutions Ltd (2010) has already evaluated the necessary data for the Kapiti Coast.
2. Estimate the range of mean sea level elevations for the time each storm occurs. Generally, the most recent IPCC projections are used, as they should represent a complete review of the available projections. Note that it is not appropriate to select either the worst case, or best case, scenarios.
3. For each storm estimate the amount of coastal erosion. This is best based on historical observations, but can be estimated by model predictions. There must be allowance for shoreline recovery between storm events, which is best determined from historical observations. Note that this model can be applied to an accreting coast by adjusting the recovery phase to incorporate the long-term trend.
4. Estimate the final shoreline position at the end of the prediction period by temporally averaging the last 2 years (this reduces the influence of any storms that occur in the last 2 years, and therefore haven't had sufficient time for the recovery phase).
5. Subtract the initial position from the final position to estimate the shoreline change (negative values correspond to erosion).
6. Repeat steps 1-5 until the exceedance probabilities > 0.01% converge (bootstrapping).

Ranasinghe *et al* (2012) found that using this approach, with the numerical SBEACH estimating the coastal erosion, and an assumed sea level rise of 0.92 m relative to 1990 by 2100, the BR method (used by CSL, 2008a, 2012) estimates corresponded to probabilities of exceedance between 8% and <1% depending on the shoreline slope used (higher probabilities associated with steeper slopes). They used BR sea level multipliers of 34-68 *cf.* 28-57 for most of the sites analysed by CSL (2008a). However, they didn't use the technique to hindcast the observed shoreline response to historic sea level rise, so it is difficult to assess how reliable the method is for forecasting.

An important aspect of the methodology suggested by Ranasinghe *et al* (2012) is recognition of shoreline recovery following storm events. This would facilitate consideration of the impacts of coastal management. de Lange *et al* (1997) developed a similar methodology to assess the overall impact of climate change on the New Zealand coast. This was extended to islands in the Pacific (Kench and Cowell, 1996), and is incorporated into the Sim-

CLIM climate impact modelling software. Based on this approach, Warrick (2006) determined that for the IPCC 2001 worst-case scenario, an annual accretion rate of 0.015 $t.m^{-1}$ of beach length would be sufficient to offset the predicted erosion. This is several orders of magnitude smaller than the observed rate of accretion for the Kapiti Coast, and suggests that the proposal of Gibb (1978) to utilise the offshore sand resource to renourish the Paekakariki to Raumati shoreline would be a successful strategy.

There does not appear to be an existing probabilistic model for predicting future inlet response. Development of a model for the Kapiti Coast will be complicated by the long history of inlet modification.

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