FIELDTRIP 5

COASTAL HAZARDS OF THE BAY OF PLENTY

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INTRODUCTION

This field trip will examine the advances in our understanding of coastal processes and hazards in the context of coastal development in the Bay of Plenty. If time permits, we will visit Bryans Beach, Ohiwa Spit, Ohope Spit, Thornton, Matata, Maketu, Papamoa, Omanu and Mount Maunganui.

COASTAL HAZARD ZONES

Bruun (1964) defined a development setback line as "an established survey line indicating the limits for certain types of developments" for the purposes of dealing with coastal erosion in Florida. This setback was determined by combining technical, developmental and administrative aspects relating to specific sites. Gibb (1981) subsequently introduced the concept of a coastal hazard zone (CHZ), where a CHZ was defined as "an adequate width of land between any development and the beach".

In practice, the first Coastal Hazard Zones (CHZs) in New Zealand corresponded to the 1 chain (~20 m) zone of riparian rights (Queen's Chain) that extended landward of the mean high water mark. However, it became clear that this distance was insufficient to provide adequate protection from large storms. After examining the maximum shoreline retreat caused by individual storms, in 1972 the Ministry of Works and Development recommended a standard CHZ of 60 m for the whole New Zealand coast (Stuart, 1984).

For some problem areas the application of a standard 60 m CHZ did not provide sufficient protection. Instead, it was necessary to develop CHZs that were site specific and involved greater widths of the coastal land. A variety of different methodologies were used to define these zones. To define a CHZ, several major processes must be assessed including (Gibb, 1981, 1987; Healy, 1985, 1991; Healy and Dean, 2000; Kay et al., 1993; Stuart, 1984):

- Geological characteristics of the site, including tectonics, structural controls and mass movement;
- Geomorphology of the beach, including the extent and character of the dunes;
- Sediment budget, including sources and sinks for sediment;
- Historical erosion rates, which may be divided into short-term changes (due either to single storms or averaged over 10 years), and long term changes (typically 100 years);
- Relative and absolute sea level changes;
- Inundation height and extent, including the effects of wave runup, storm surge and tsunami;
- River mouth or tidal inlet mobility, for sites located on barrier spits; and
- Presence and effectiveness of protection works.

Not all processes were formally included in every method used. The methods also differed in the weighting applied to the contributions of the processes included, and the mechanics of how they are combined. They also tend to require a degree of subjectivity in their application, so are semi-quantitative.

It should be noted that the only consideration of anthropogenic factors is in relation to the mitigating effects of coastal protection works. The CHZ strictly does not include the presence of infrastructure, public perceptions, natural character, or cultural concerns, although these may be factors to consider under the RMA (1991). Further, strictly the CHZ should consider all natural hazards with some probability of occurrence, which may include extremely rare events that may or may not have a significant effect (bolide impact for example). In practice, the hazards considered are constrained by the inclusion of a planning horizon leading to a minimum annual exceedence probability (AEP). Typically

AEPs are in the range 1-3%, although more extreme values are occasionally used, such as 0.0001% for coastal defences in The Netherlands.

For New Zealand, the main hazards considered are typically:

- Beach and cliff erosion;
- Storm tide inundation – the combined effect of storm surges and astronomical tides;
- Dune blowouts and transgressive sand sheets; and
- Tsunamis.

In some areas, the effects of intense rainfall should also be included, particularly where coastal development has modified the catchment and drainage characteristics. Some catchments are prone to episodic debris avalanches, and unfortunately the alluvial fans created by previous events have tended to be primary sites for coastal development along cliffed coasts. It may also be appropriate to consider tectonic effects depending on the planning timescale being considered.

DEVELOPMENTS IN UNDERSTANDING

Our understanding of the processes that create and mitigate coastal hazards is generally improving over time. One difficulty that arises is the incorporation of that improved understanding into coastal management. For example, we have much better estimates of sea-level response to global warming in 2007 than we had in 1990, yet the 1990 estimates are often considered the most appropriate for coastal management because they are "more widely accepted" (Healy and Dean, 2000).

This field trip will consider changes in our understanding for some aspects of coastal hazards and see the potential impacts of these on the Bay of Plenty coast. The particular aspects considered are:

- Beach responses to wave forcing (morphodynamics);
- Beach response to water level changes, particularly due to storm surge and storm tide, and sea level rise;
- Climatic variability and its influence on the coast; and
- Tectonic effects.

Beach morphodynamics

During the last two decades a growing focus of coastal studies has been on understanding the dynamic equilibrium between physical processes and the changing shoreline morphology (beach morphodynamics). This has led to an improved understanding of the relative importance of factors that influence the response of a beach to storms. The main factors that control the morphodynamic response of a beach are the beach slope (which is related to sediment texture) and wave steepness (which depends on wave height and period). Combining these parameters it is possible to recognise a range of beach states and their associated forcing conditions (Wright and Short, 1983), and hence predict likely beach response. To facilitate prediction, wave steepness and beach slope are combined to form a non-dimensional surf similarity parameter, where beaches with similar morphodynamic characteristics have roughly equal values of the surf similarity parameter.

Two surf similarity parameters are useful predictors of beach response for New Zealand. These are defined in terms of breaking root mean square wave height (H_b), deep water wavelength (L) and beach slope (m):

- Nearshore or breaking Iribarren number ξ_b, which may be expressed as (Battjes, 1975)

$$\xi_b = \frac{m}{\sqrt{H_b / L_\infty}}$$

- Dean's Parameter Ω_b (Dean, 1973), which uses wave period (T) as a surrogate for wavelength, and sediment setting velocity (w) as a surrogate for beach slope. It is defined by

$$\Omega_b = \frac{H_b}{wT}$$

Various schemes have been proposed to classify the range of possible beach states. Three main beach states have been recognised (Komar, 1998):

- *Dissipative* ($\varepsilon<2.5$, $\xi_b>1.125$, or $\Omega<1$) — flat beach profiles where the initial breaker zone is well offshore. The waves travel inshore as well developed bores and may reform to form a succession of breaker zones and surf zones. As the offshore wave height increases, the initial breaker zone moves further offshore so that there is little change in the wave height at the shore. Hence this type of beach is very efficient at dissipating wave energy.
- *Reflective* ($\varepsilon>20$, $\xi_b<0.34$, or $\Omega>6$) — steep beach where the initial breaker zone is very close to the beach. There is usually only one breaker zone, and the swash motions are strong. An increase in the offshore wave height is associated with a corresponding increase at the shore. This type of beach does not dissipate much wave energy, and much is reflected offshore.
- *Intermediate* — these lie between the previous two types in terms of their behaviour. They are characterised by complex three-dimensional morphologies that are associated with complex water circulation systems. Due to the complex morphologies, it is possible to subdivide intermediate beaches into a range of different sub-types.

Several important generalisations need to be considered when evaluating the beach state (Wright and Short, 1983, 1984):

- the beach state may be dissipative, reflective, or in some intermediate state depending on local environmental conditions, sediments, and antecedent wave conditions;
- the relative contributions of incident waves, infragravity waves, and net surf zone circulations to near-bottom currents, and hence the resultant sand transport, vary with beach state;
- the actual processes that cause beach cut, and the wave energy required to induce beach cut, are dependent on beach state;
- as beach state changes in response to changing morphology or hydrodynamic processes, the influence of various hydrodynamic processes varies, and the beach state may become (temporarily) independent of deep water conditions; and
- the normal range of temporal change exhibited by the beach and surf zone shows a close relationship to the most common beach state.

Given the above it is possible to formulate a model for beach state. One such model, the Wright and Short model, consists of six beach states that vary between the dissipative and reflective extremes (Wright and Short, 1983, 1984).

However, prediction is complicated by the antecedent conditions, because the response of a beach to forcing conditions depends on the initial state that it is in (Horikawa, 1988). Further, the speed at which equilibrium with the new forcing conditions varies depending on whether wave energy is increasing or decreasing. Therefore, a beach generally responds more quickly to storm conditions than to fair weather conditions.

This approach does not explicitly include the effect of water level. It is recognised that a beach may exhibit different beach states at different tidal heights (Horn, 1993; Massel et al., 1993). Further, numerical simulations of beach profile evolution indicate that water level is 10-20 times more significant that wave height in determining the extent of subaerial erosion (Kreibel et al., 1991).

Erosion/accretion criteria

Simple parameters have been developed that can be used to predict the net direction of diabathic sediment transport within the nearshore zone. Net onshore movement is taken to indicate beach accretion, and net offshore movement is taken to indicate beach erosion. These relationships are often used to define transitions in beach state models, since the changes in beach state tend to be associated with a redistribution of sediment within the beach profile. The following relationships are simple criteria for predicting whether a beach will erode or accrete by wave-induced diabathic sediment transport. These criteria ignore any longshore sediment transport that may occur, and assume that sufficient sand is present to allow erosion or accretion.

A large number of non-dimensional parameters have been suggested. Of these, the Deans parameter based on the deep water significant wave height (Ω_∞) appears to be the best single parameter predictor, with a critical value of 3.2. This single boundary model can be further refined to give (Kraus et al., 1991):

If $\Omega_\infty < 2.4$, then accretion is highly probable.

If $\Omega_\infty < 3.2$, then accretion is probable.

If $\Omega_\infty \geq 3.2$, then erosion is probable.

If $\Omega_\infty > 4.0$, then erosion is highly probable.

Hallermeier limits

Hallermeier limits (Hallermeier, 1980, 1981) can be invoked to partially explain this behaviour. These limits define the offshore extent of active sediment transport under storm conditions (Inner Limit), and the maximum offshore extent of sediment movement under average wave conditions (Outer Limit). During a storm sediment is transported offshore. As the storm wanes, sediment landward of the Inner Limit can be easily transported back to the beach. However, sediment transported to between the Inner and Outer Limits (shoal zone) moves only very slowly. Data for Tauranga (de Lange and Healy, 1994), Pakiri Beach and Australian east coast beaches (Hesp and Hilton, 1996) indicate that the return of sediment from this zone may take decades. Sediment transported beyond the Outer Limit is permanently lost from the system.

Hence, Hallermeier limits provide a physical explanation of the movement of sediment. However, to fully explain the observed pattern of shoreline erosion and accretion it is also necessary to change the temporal distribution of storm events. Clearly if storms are sufficiently close together, sediment will progressively be displaced from the beach into the shoaling zone and possibly beyond the Outer Limit. Alternatively, if the storms are sufficiently far apart, sediment in the shoal zone can be returned to the beach producing stability.

Sediment availability

Associated with the concept of Hallermeier Limits is that of sediment availability. Based on work at Pakiri (Bell et al., 1996), and at Mt Maunganui (de Lange and Healy, 1994), it is clear that the Hallermeier Limits are good indicators of zones of potential sediment movement for the New Zealand coast. However, this is obviously of little consequence if suitable sediment is unavailable.

Many beaches along the Bay of Plenty coast experienced erosion between 1950 and 1980, and this included the section between Mt Maunganui and Papamoa.

Table 1. Summary of mid-late Holocene progradation rates for four coastal plains on the east coast of the North Island

Location	Study	Rate ($m.y^{-1}$)*
Papamoa	Wigley (1990)	0.2
Rangitaiki Plains	Pullar & Selby (1971)	0.5-0.7
Whitianga	Abrahamson (1987)	0.5
Gisborne	Pullar & Penhale (1970)	0.5-0.8

* ^{14}C years

However, erosion at Ocean Beach, Mt Maunganui, was less severe and the beach appeared to recover sooner than adjacent stretches of coast. Early studies into the dispersal of dredge material off Mt Maunganui suggested that some of the sediment may have been transported shoreward (Dahm and Healy, 1980). This was subsequently confirmed by further studies of the dump grounds (Harms, 1989; Healy et al., 1991; Warren, 1992). Hallermeier Limits were found to provide a useful prediction of the likely behaviour of the dredge spoil (Hands and Allison, 1991), and these were used to deliberately use dredge spoil to renourish Ocean Beach (Foster, 1991; Foster et al., 1996). This work confirmed that dredge spoil could be used to increase sediment availability and enhance beaches. Further, it was possible to vary the depth of emplacement to control the rate at which the spoil renourished the subaerial beach. The magnitude of the response to the addition of dredge spoil suggests that modern beaches in the Bay of Plenty have a limited sediment availability that may affect their response to coastal processes.

Various studies have examined the development of coastal plains around New Zealand. Table 1 summarises the average progradation rates determined by a small selection of these. Not evident from this table is the temporal variability of the progradation rates. At most sites there is considerable variation of time, and the question arises whether these rates have any significance for the definition of a coastal hazard zone.

At many sites there appears to be a period of rapid accretion starting around 6500-6000 ^{14}C BP when sea level reached levels similar to those of the present. Subsequently, rates have generally slowed. Table 2 summarises the variations in progradation rates determined for Papamoa by Wigley (1990). These data indicate an overall rate of 0.215 $m.y^{-1}$ for 14C dates, or 0.188 $m.y^{-1}$ for revised calibrated years, which is low, compared to rates for other coastal plains on the east coast of the North Island.

The rates at Papamoa (Table 2) have been reassessed in light of additional tephra identification (Alloway et al., 1994; Dahm et al., 1994; Lowe et al., 1992; Newnham et al., 1995). These studies concluded:

- the coastal plain began to develop between 5600-6300 ^{14}C BP, but initially the extent was limited to a single, wide dune;
- between 4085-4530 ^{14}C BP there was a rapid advance coincident with major volcanic eruptions;
- this was followed by a relatively high rate of progradation until ~3500 ^{14}C BP; and
- the last 700-1000 years are characterised by very low rates of progradation.

Table 2. Summary of progradation rates at Papamoa. The rates (revised here) were originally determined by Wigley (1990) using marker tephra layers preserved in shore parallel coastal ridges and peats. Ages on tephras are from calibration of Stuiver *et al* (1998) and from Lowe and de Lange (2000) and Hogg *et al* (2003). Tephra names are from Froggatt and Lowe (1990), with alternatives in parentheses from Wilson (1993).

Progradation rates during Late Holocene at Papamoa

Shoreline	Marker Tephra	Age (cal y BP)	Distance (m)	Progradation (m)	Time (y)	Rate ($m.y^{-1}$)
Start		7250	0			
				80	2050	0.040
R2	Hinemaiaia (Unit K)	5200	80			
				730	2400	0..304
R16	Whakaipo (Unit V)	2800	810			
				100	1050	0.095
R18	Taupo (Unit Y)	1750*	910			
				350	1115	0.314
R23	Kaharoa	635**	1260			
				68	571	0.119
R24	Tarawera 1886	64	1328			
				40	104	0.385
1990		-40	1368			

Ages are calibrated (calendar) years before 1950 AD. *c. 200 AD ** c. 1315 AD

These data suggest that, although Papamoa is a cuspate foreland, coastal progradation is probably limited by the available sediment supply. However, episodic volcanic eruptions and land-use changes (Giles et al., 1999; Lowe et al., 2000) may boost the available sediment, resulting in a short period of significant accretion. One mechanism recently identified as supplying a large amount of sediment to the coast, is a so-called “breakout flood” following a volcanic eruption.

For example, in New Zealand during the Holocene, sediment supply to the coast by the Waikato and Tarawera rivers has been strongly modified by breakout floods from Lakes Taupo and Tarawera respectively (Hodgson and Nairn, 2005; Manville et al., 1999; White et al., 1997). Following the 1886 Tarawera eruption there appears to have been little initial erosion of tephra deposits in the Tarawera River catchment, so that the only eruptive material reaching the coast was that supplied directly as fallout tephra. In November 1904, a debris dam created during the eruption near the outlet of Lake Tarawera failed, causing a two-day long breakout flood. This was followed by several decades of intense erosion in the upper catchment, and aggradation in the lower catchment and coast (White et al., 1997). The breakout flood was caused by a 2 m increase in the elevation of Lake Tarawera.

The earlier Kaharoa eruption in c. 1315 AD produced at least a 30 m increase in lake level, and a correspondingly larger breakout flood event (Andrews, 1999; Hodgson et al., in prep). This breakout flood was associated with a marked advance of the coastline of the Rangitaiki Plain between Matata and Whakatane (Pullar and Selby, 1971).

The availability of extra sediment following an eruption could therefore be seen as reducing coastal hazard, because of increased coastal accretion. However, apart from the hazard directly posed by the breakout floods on coastal plains, the temporary increase in progradation rates can distort assessments of coastal hazard zones. In the case of communities on the river channels and along the coast, such as Papamoa, the rate of progradation for the last century is almost twice the long-term trend. If this is due largely to the 1886 Tarawera eruption supplying additional sediment, then progradation should be slowing down.

This interpretation may be distorted by the use of tephra to date the beach ridges, since this tends to highlight volcanic eruptions. There may be other variations in sediment supply that affect the coast, such as climatic variability and land use changes. These are considered below.

Influence of water level

An intuitive conclusion would be that increasing wave height results in greater erosion. This often occurs, leading to the further conclusion that wave height is the major control on coastal erosion, and hence areas with lower wave heights should have less erosion. Unfortunately, this is not the case. For example, wave steepness is more important than wave height, and so large low steepness waves can result in accretion as often happens as a storm wanes. This is evident from the beach state models and erosion/accretion criteria discussed above.

A combination of numerical and physical modelling with detailed observations of beach morphodynamics have demonstrated that water level is the major factor affecting coastal erosion. Part of the increased erosion observed with larger waves is due to the higher water level they produce through wave set-up. This is enhanced by long groups of large waves. Water level has been demonstrated to be 10-20 times more important than wave height, with the average value being around 16 times (Kreibel et al., 1991).

The relationship between water level and erosion is most significant if the waves reach the boundary between the beach and the foredune (Allan *et al*., 2003). If this occurs, then the criteria for erosion/accretion discussed above no longer appear to be valid, and erosion is highly likely regardless of the wave steepness. It is probable that different processes are involved once the wave swash reaches the foredune. This has led to a proposal that the overall erosion potential of a storm should be defined by the duration of water levels above the elevation of the foredune-beach boundary (Ruggiero et al., 2001; Zhang et al., 2001). Such measurements may be calibrated against historical erosion events to provide an assessment of potential erosion for future storms, provided the height and duration of storm water levels can be predicted.

The water level during a storm is a combination of the wave set-up and the storm tide. In practice, it is usually more useful to consider the wave contribution to increased water levels in terms of the wave run-up. Run-up is the maximum swash elevation above the still water level, and so it combines wave set-up and the swash excursions. Use of wave run-up allows consideration of variability of wave heights during a storm. The storm tide is the combination of the astronomical tide and storm surge due to reduced atmospheric pressure and wind stress.

Over longer time periods, changes in mean sea level also affect coastal erosion, and normally should be considered in determining coastal hazard zones. The media frequently highlights concerns about the consequences of sea level rise resulting from Greenhouse Gas driven enhanced global warming. Unfortunately, this does tend to reduce awareness of the much larger natural variations that occur over shorter time scales.

Storm tides

A storm tide is the increased water level resulting from the combination of a high tide and a storm surge. For defining coastal hazards, storm tides are a better measure of extreme water levels than storm surges. A large storm surge occurring a low tide may represent a smaller hazard, than a smaller event coinciding with the largest spring tide of the year. This was demonstrated by the sequence of storm surges associated with Cyclones Fergus, Drena and Gavin in the summer of 1996/97. Of these the largest storm surge was associated with Cyclone Gavin, but it had negligible effect. The smaller surges caused by Fergus and Drena were associated with higher tides and did considerably more damage.

Approaching the problem in terms of storm tide elevations allows predictions to be made of future periods when the hazard is more likely to occur. The largest storm tides are most likely to occur in associated with the largest high tide levels. These correspond to perigean spring tides: during a full or new moon lunar phase when the Moon is closest to the Earth. Perigean tides can be predicted from lunar orbital characteristics without know the details of tidal wave behaviour for any location of interest.

Storm tides create coastal hazards by increasing coastal erosion and also by inundation. As discussed above, water level is typically 16 times more important than wave height in determining the extent of erosion during storm conditions. This appears to particularly important if the increase in water level allows waves to extend landward of the toe of the foredune. The risk of coastal erosion may be expressed as the number of hours per year that the water level permits wave excursions landward of the foredune toe (Ruggiero et al., 2001). An important component of this determination is therefore the distribution of storm tides.

Goring *et al* (1997) have undertaken an analysis of extreme water levels on the Bay of Plenty coast using data obtained at Moturiki Island (the longest open coast water level record available in New Zealand). These data could be used to predict the erosion risk. However, two difficulties have been identified with the application of these data:

- Observed storm tide levels around the Bay of Plenty following cyclones Fergus and Drena differed significantly from those measured at Moturiki Island (Blackwood, 1997);
- A longer time-series of storm surge measurements from within Tauranga Harbour indicate that storm surge frequency and magnitude has varied on decadal scales, and the Moturiki data may not reflect the true long-term distribution of extreme water levels (de Lange and Gibb, 2000b).

Considering the observed response during cyclones Fergus and Drena, Blackwood (1997) concluded that the Moturiki recorder under-reported the storm surge by around 0.5 m, and hence expected storm-surge hazard elevations in the Bay of Plenty were at least 0.5 m too low. At least part of the discrepancy arises due to the location of the instruments and observations used in the analysis. The Moturiki data were obtained from an open coast tide gauge, located on the seaward side of an island connected to the shore by a small cuspate foreland and short tombolo. The remaining observations were obtained within estuaries at Whakatane, Ohiwa and Opotiki, and were taken as extreme water levels. The extreme water level at those sites include variations in tide level not necessarily accounted for in the secondary port tidal corrections, any thermosteric responses to the storm effects in the estuary catchments, wave set-up, and local wind effects within the estuary. All these effects result in higher water levels inside the estuary compared to the open coast (de Lange and Gibb, 2000b; Gibb, 1997). Therefore, extreme water level distributions should be considered to be site specific.

The decadal-scale variability in storm surge behaviour reported by de Lange and Gibb (2000b), arises from decadal scale climatic variations in the forcing processes. These will be discussed below. However, one consequence of these variations that has yet to be addresses in CHZ determinations is that the AEP distributions over time-scales shorter than these natural variations may vary considerably. This can result in CHZ determinations being too conservative (which may be acceptable), or too optimistic (which may not).

Sea level rise

The location of mean sea level at the coast is a function of global sea level, regional oceanographic (eg. tides and ocean currents) and atmospheric processes (eg. barometric pressure and winds), and land movements (Pugh, 2004). Global sea level has been rising since the end of the Little Ice Age, albeit with regional variations. New Zealand has experienced this rise, and all available evidence indicates that this is likely to continue for the immediate future.

Assessing sea level rise for New Zealand has been difficult due to the shortage of suitable measurements. Most records have been obtained from ports and they are generally of poor quality (Goring and Bell, 1996). Bell *et al* (2000) reviewed the available data and summarised the behaviour of sea level over the 20th Century. Extracting the local sea level rise is complicated by large interannual variations. Different rates can be obtained by selecting different periods of time (Holgate, 2007).

The most extensive analyses of seasonal to decadal-scale sea level fluctuations for New Zealand are based on sea level measurements around Tauranga (Bell and Goring, 1998; Bell et al., 1999; de Lange and Gibb, 2000b). These results indicate the likely variability of sea level at Ohope, and have identified the main contributions to annual mean sea level variability (excluding land movements) as:

- Seasonal variations due to heating and cooling of coastal waters, and changes in atmospheric conditions (±80 mm);
- ENSO variations due to changes in ocean circulation, changes in atmospheric conditions and heating and cooling of surrounding oceans (±120 mm);
- IPO variations affecting the magnitude and frequency of ENSO events, and distribution and paths of storms (±50 mm);
- Global sea level changes.

It has also been suggested that sea level records elsewhere display longer-term oscillations (Fairbridge, 1989, 1998). The New Zealand data are not suitable to assess the validity of these assertions.

Hence, the year-to-year elevation of mean sea level in the Bay of Plenty may vary by ±250 mm. Over short to medium time scales (10-50 years) this variability is more important than sea level rise.

Most sea level rise predictions for AD 2050 and AD 2100 assume acceleration in the rate of sea level rise over the late 20th Century and throughout the 21st Century (viz. Table 3). An acceleration was not observed for New Zealand during the 20th Century (Hannah, 2004). Careful analysis of global sea levels as reported in the Third Assessment Report (IPCC, 2001) also did not detect any acceleration. The Forth Assessment Report (IPCC, 2007) suggests that acceleration may be occurring by comparing short periods of tide gauge and satellite data.

However, analysis of long-term sea level data indicates that the IPCC (2007) results merely reflect decadal variability (Holgate, 2007). Reviews of global sea level records have recently demonstrated that sea level rise is slowing, and noted a lack of any significant predictive relationship between global temperature and sea level since the 1600s (*viz*. Larsen and Clark, 2006; Holgate, 2007).

Table 3 – Predictions of expected sea level rise by AD 2100 relative to AD 1990 levels from various sources. Note that the predicted level has decreased with time, and that the Forth Assessment Report (IPCC, 2007) utilised a different method to evaluate sea level that is not directly comparable to all the other studies.

	Sea level rise by 2100 (mm)		
Source	Minimum	Maximum	Most likely
Hoffman *et al* ((1983) - US EPA report	1440	2170	1805
Hoffman et al (1986) Thomas	600 600	3700 2300	2150 1100
IPCC 1990	310	1100	660
IPCC 1995	200	860	490
IPCC 2001 – most likely - extremes	310 90	490 880	440 485
Suzuki *et al* (2005)	-90	300	230
Church and White (2006)	280	340	310
IPCC 2007 (*estimated most likely*)	180	590	385
Historic NZ rate	130	170	140

Predicted future sea levels are dependent on accelerating rates of sea level rise during the 21st Century. There are highly variable predictions of the extent of global warming and this impacts on the resultant sea level changes. The absence of acceleration during the late 20th Century and changing understanding of the response rates of processes that contribute to sea level rise have resulted in a reduction in the predicted change by AD 2100 with successive reassessments (Table 3). One study based on sea level data from 1870 to 2004 has reported an acceleration of sea level in the first half of the 20th Century (Church and White, 2006), which results in a higher sea level than predicted by the majority of computer models (viz. Suzuki et al., 2005). However, all recent studies have identified that decadal scale processes have a much greater influence on coastal sea levels than global sea level, as noted for New Zealand by Bell *et al* (2000).

One aspect of sea level rise impact that has not progressed over the last 20 years has been the development of methodologies to assess the impact of sea level rise on shoreline position. The most widely applied method for CHZ determinations has been the “Bruun Rule” (Bruun, 1962, 1964; Healy and Dean, 2000). Despite sophisticated diagrams used to justify the application of this Rule, it basically states that there is a linear landward translation of the shoreline defined by the amount of sea level rise and the slope of the beach, as given generally by:

$$\text{shoreline retreat} = \frac{\text{sea level rise}}{\text{beach slope}}$$

Different methods can be applied to define the slope of the beach, which produce a range of shoreline retreats. Ignoring the relative merits of the different slope determinations, there are many problems with applying such a simple 2-dimensional relationship to a complex coast (Bruun, 1983, 1988; SCOR Working Group 89, 1991). The use of the Bruun Rule should be tempered with consideration of historical trends for any site, especially given at least 100 years of historic sea level rise for the New Zealand coast.

The Bruun Rule can be expressed mathematically as

$$R = K \times S$$

Where R = new location of shoreline (m), which is often assumed to involve erosion; S = sea level rise (m); and K = scale factor equal to the reciprocal of the slope.

Zhang *et al* (2004) evaluated K for the Florida coast based on historic sea level rise and shoreline change. They avoided any sites subject to human activities such as sand mining or close to tidal inlets, and they found values of K ranging from 50-120, with a mean of 78.

A similar analysis has not been published for the Bay of Plenty, although the data are available. Tonkin and Taylor (2002) analysed the historical response to sea level rise for Ohope Spit, and their data can be used to evaluate K. Assuming an average annual sea level rise rate of 1.6 mm.y^{-1} for the period of 1868-1997, historical K values are given in Table 4. There is considerable variation in values, but they indicate that the western end of Ohope has accreted while sea level has risen, while the coast east of approximately Maraetotara Stream has eroded. The historical values also display the opposite trend to the underlying assumption in the Bruun Rule, which would predict the greatest erosion at the western end of Ohope Beach.

Table 4 – K values (Equation 2) determined for historic shoreline change at Ohope between 1868-1997 assuming a sea level rise of 208 mm (slightly higher than the 195 mm global rise for 1870-2004 reported by Church and White (2006)). Note that negative erosion corresponds to progradation. The assumed K values are those used by Tonkin and Taylor (2002) to evaluate the SL term.

Site	Shoreline erosion (m)	K value	Assumed K
No. 63 West End	-121.5	-584	78-100
No. 31 West End	-65.9	-316	78-100
Surf Club, Mahy Stream	-39.1	-188	78-100
Maraetotara Stream	10.1	49	68-81
Moana Rd	29.0	139	68-81

The SL term is incorporated into Equation 1 to account for the effects of accelerated sea level rise: so the purpose is to account for the possibility that the as sea level rise increases the pattern of shoreline change will change. The difficulty is that the Bruun Rule can only predict shoreline erosion, because in application it does not accommodate negative K values (Bruun, 1983; Pilkey and Cooper, 2004). Low and negative K values arise due to factors such as longshore sediment transport dominating over the effects of sea level rise, and an increased beach slope due to accumulation of sand in dunes. Dune restoration would mitigate sea level rise effects as predicted by the Bruun Rule, by increasing the beach slope (reducing K values) resulting in reduced erosion.

For the Bay of Plenty, this is demonstrated by the response of the coast west of Whakatane following the 1987 Edgecumbe Earthquake (Ruscoe, 1988). Subsidence associated with the event resulted in a 400 mm relative rise in sea level at Thornton Beach (comparable to the predicted AD 2100 rise based on IPCC 2007). In response, the shoreline should have eroded by ~15 m. This did not occur. Instead the shoreline has continued to prograde, and shoreline changes at Thornton continue to be dominated by ENSO extremes (Pickett et al., 1997).

Tectonic effects

The Bay of Plenty coast is tectonically active, so the coast may undergo uplift or subsidence (Beanland and Berryman, 1992). Ohope Spit and Ohiwa Harbour demonstrate some of the impacts of uplift and subsidence on coastal stability.

Ohiwa Harbour is a drowned river system eroded into Pleistocene sediment rocks between the Ohope and Waikaremoana Faults. The river mouth has been partially blocked by the

Holocene Ohope and Ohiwa Spits. It has been suggested that the extensive intertidal flats (70% of the harbour area) are due to extensive sedimentation during the Holocene, with the sedimentation mostly occurring during the last 2 thousand years once infilling of the Whakatane Graben permitted bypassing of Whakatane Head (Hayward *et al*., 2004; Richmond *et al*., 1984). It is assumed that the rivers entering Ohiwa Harbour do not supply much sediment.

It has been long recognised that the hills around Ohiwa Harbour are capped with shallow marine, sedimentary rocks (Ohope Formation) that used to form a planar surface tilting down towards the east from >60m elevation next to the Whakatane Fault to ~10 m near the Waikaremoana Fault (Hayward et al., 2004). The elevations of the marine beds and other sea level indicators have been used to determine rates of uplift (Beanland and Berryman, 1992; Pillans, 1986, 1990). The rates determined by different authors vary (Murdoch, 2005), but consistently show the highest rate at the Whakatane Fault (~1 $mm.y^{-1}$), and decrease eastwards to a minimum at the Waikaremoana Fault (~0.2 $mm.y^{-1}$).

The uplifted planar surface is fragmented into four main fault blocks, with Ohope Beach and Ohiwa Harbour straddling three of these. The Waiotahi Estuary occupies the fault-angle depression on the western margin of the fourth block (Marra, 1997). In addition to differential uplift across the faults bounding the blocks, which causes tilting of their surfaces, there is horizontal displacement along the faults of 0.7-1.1 mm/y^{-1} (Beanland and Berryman, 1992), which can offset western sides of the shoreline of Ohope Beach and Spit where the faults interest the coast.

In addition to uplifted marine beds, there is also clear evidence of subsidence that forms fault angle depressions on the eastern margins of the Ohope, Waimana and Waikaremoana Faults. The Waimana Fault is linked to two areas of submerged coastal forest within the harbour (Murdoch, 2005; Pullar and Patel, 1972) and the active beach face along Ohope Spit (Julian, 2006). These are linked to two seismic subsidence events. The subsidence rates have not been determined, but are likely to be similar to those in the Whakatane Graben, which is subsiding at 1-2 $mm.y^{-1}$ (Beanland and Berryman, 1992). Subsidence occurs episodically in association with earthquakes. Three events have been recognised for Ohiwa Harbour, but a detailed analysis of recurrence time has not been done. As an indication, the recurrence of subsidence causing earthquakes for the Whakatane Graben is 200-300 y, with the last event in 1987 (Beanland and Berryman, 1992).

Therefore, parts of Ohope Beach and Spit are being uplifted, while other parts are subsiding. The work of Murdoch (2005) suggests that Ohiwa Harbour occupies a subsiding basin whose extent has increased following each major subsidence event. This contrasts with previously published studies.

Richmond *et al* (1984), postulated that Ohope Spit started to form around 4000 BC with the eastward extension a spit driven by eastward littoral drift around Whakatane Point. This growth was matched by a gradual retreat of a pre-existing Ohiwa Spit. Dating of the evolution of the spit was based primarily on dated tree stumps and tephras mapped by Pullar and Patel (1972) and Gibb (1977). Adoption of this model implies that stability of Ohope Beach and Spit is dependant on the supply of sediment from the Whakatane River, and ignores the effects of subsidence and uplift.

Murdoch (2005) undertook a comprehensive review of all available data for soil development on Ohope Spit and within Ohiwa Harbour, and supplemented this with a field mapping and coring programme. His data indicate that the Ohope Spit is of similar age to all the other spits and coastal barriers mapped around the coast of the Bay of Plenty and eastern Coromandel Peninsula, such as the Papamoa cuspate foreland.

Progradation rates are not as well preserved for Ohope Beach and Spit as they are at Papamoa. However, the available data indicate an average of 0.27 $m.y^{-1}$ from 1718 to 636 cal. yrs BP, reducing to 0.06 $m.y^{-1}$ from 636 cal. yrs BP to the present. Observations reported by Saunders (1999) and Julian (2006) indicate that currently the seaward progradation rate is essentially zero, possibly due to the reduction of sediment supply due to activities such as sand and gravel extraction from the Whakatane River (Eynon-Richards, 1988).

Although the development of Ohope Spit is similar to the general pattern for the Bay of Plenty (initially forming ~7200 cal. yrs BP; rapid expansion ~5000 cal. yrs BP; decreasing rate of progradation until stable state over last millennium), there are important differences. These are mostly linked to influxes of sediment associated with volcanic eruptions from the Taupo Volcanic Zone. The key phases for Ohope spit are:

- Slow initial progradation as sea level rose towards present levels from ~7200 to 5590 cal. yrs BP. At this stage a thin beach existed at West End and extended eastward to form a short spit terminating around Liddon Cove or Elizabeth St.
- Following the eruption of the Whakatane Tephra, Ohope Spit prograded eastward rapidly, reaching to within <400 m of the AD 2000 terminus by 4190 cal. yrs BP. This rapid expansion coincided with the onset of the present sea level still-stand and corresponds to the rapid expansion of other barriers around the Bay of Plenty. Growth of Ohope spit appears to have been faster than other spits, and this is attributed to influx of volcanic sediment following the eruption of the Whakatane Tephra.
- From 4190 to 2780 cal. yrs BP the spit continued to accumulate sediment, growing wider and higher with the development of a major dune along the length of the spit. The harbour entrance appears to have been maintained in the present position since the end of this period of development, and it is suggested that the location is constrained by a river valley formed when sea level was lower during the previous glacial.
- Around 2600 cal. yrs BP an earthquake on the Waimana Fault is inferred to have caused 2.0-2.4 m of subsidence of the eastern side of Ohiwa Harbour. Associated with this event, the spit was breached between Phillip and Anne Streets. Stumps of totara trees dating back to this period are evident in the beach east of Anne Street. It is presumed that these were killed by inundation following subsidence, and that the spit extended further seaward at that time. It is unclear whether the spit breach was a permanent second entrance, or only opened intermittently.
- The spit began another phase of rapid growth following the Taupo Eruption (1718 cal. yrs BP), which resulted in a seaward expansion of the spit and complete closure of the spit breach. The Whakatane Graben to the west appears to have been full of sediment at this time, with the coast close to its' present position west of Whakatane, so additional sediment reached Ohope around Whakatane Head. The increased volume of sediment resulted in the formation of a series of shore-parallel dune ridges, and infilling of Ohiwa Harbour.
- Between 1718 and 1110 cal. yrs BP another earthquake, probably on the Waimana Fault caused subsidence of Ohiwa Harbour and most of Ohope Spit. The amount of subsidence is unclear, but is of the order 1-2 m. A submerged totara forest on the southern side of the side of the Ohope Spit is a consequence of this event.
- The Kaharoa Eruption (636 cal yrs BP) provided another influx of sediment to Ohope and Ohiwa Spits. This contributed to the extension of Ohiwa Spit into Ohiwa Harbour, where a spit breach resulted in the formation of Whangakopikopiko Island, An earthquake occurred on the Waikaremoana Fault around the time of the Kaharoa Eruption, causing subsidence at Waiotahi Estuary to the east (Hayward et al., 2004). It is not clear what affect this had at Ohiwa Harbour.
- The sea-rafted Loisels Pumice was deposited along the Ohope Spit and within Ohiwa Harbour at 590 cal. yrs BP, possibly due to an eruption at the submarine Mt Healy caldera to the northeast (Wright et al., 2003). There are two types of Loisels Pumice deposit (Wellman, 1962): a primary deposit showing little abrasion; and a secondary deposit showing evidence of reworking. Previous interpretations of coastal progradation have assumed that both deposits mark the location of the shoreline at the time of deposition. It is more likely that the primary deposits represent the inundation limit of tsunami waves associated with the eruptions producing the pumice (de Lange and Moon, 2007; Lowe and de Lange,

2000). This interpretation indicates that the rate of progradation at Ohope has decreased significantly since the Kaharoa Eruption.

Historical data have been used to evaluate changes since the 1860s. Early studies identified an progressive eastward shift in the location of the harbour entrance. However, erosion of Ohiwa Spit involved only minor movement of the harbour entrance as the spit was highly recurved. More recently, it has been recognised that there are decadal scale fluctuations in the shore-line positions of both Ohope and Ohiwa Spits, and the location of the harbour entrance (Julian, 2006; Murdoch, 2005). These fluctuations have been attributed to: the ENSO 3-7 year cycle; and the IPO 50-70 year cycle.

The mechanisms that cause erosion or accretion at any particular site are still being debated. Key factors include the frequency of flood events in the lower Whakatane River that affect the supply of sediment around Whakatane Head; variations in the mean direction of wave approach that affect the local directions of sediment transport along the spits; variations in the relative frequency of onshore and offshore winds that affect the cross-shore transport of sediment; and variations in the location and orientation of the ebb-tidal jet that affect the movement of shoals across the ebb tidal delta. One factor not yet considered is the episodic reconfiguration of the harbour entrance in response to tsunami events, particularly tsunamis from South America in 1868, 1877 and 1960 (viz. Gibb, 1977; Murdoch, 2005).

The role of climatic variability

An assessment of shoreline changes and coastal stability at Pakiri Beach, north of Leigh on the northeast coast of New Zealand, demonstrated that the beach was severely impacted by coastal erosion. This erosion became progressively worse during the 1970s, reaching a maximum extent following three closely spaced storms in 1978 (Hilton, 1990). This pattern was similar to that experienced at many beaches along North Island's northeast coast, including the Bay of Plenty. At Pakiri Beach, sand extraction was considered to be a contributing factor to the accelerated coastal erosion. However, although sand extraction continued, the beach steadily accreted over the next 20 years. A similar pattern of rapid erosion, followed by decades of accretion, was also observed elsewhere in New Zealand and Australia (Hesp and Hilton, 1996; Hilton and Hesp, 1996).

El Niño – Southern Oscillation (ENSO)

It is now recognised that New Zealand experiences changes in weather patterns over 2-7 year cycles, associated with north-south movements (Southern Oscillation) in the South Pacific Convergence Zone. These movements are related to the strength of the sea surface temperature gradient across the equatorial Pacific Ocean, and the phenomenon is known as the El Niño-Southern Oscillation (ENSO). Extremes of this oscillation are known as La Niña and El Niño in recognition of their impacts on the north-western coast of South America.

A link between ENSO extremes and the wave climate of the west coast of North America was first suggested in 1984 following a severe El Niño event in 1982-83 (Seymour, 1998; Seymour et al., 1984). The El Niño extreme is associated with larger waves, higher sea levels, more onshore winds, and increased coastal erosion on the west coast of the USA.

The New Zealand coast also experiences changes associated with ENSO extremes (Hume et al., 1992). These are best documented for the northeast coast of New Zealand, where some of the longest time series of observational data are available. An open coast tide gauge has been operating continuously at Moturiki Island, Tauranga since 1973. An analysis of the sea level record shows a strong correlation with ENSO (Goring and Bell, 1999). Sea surface temperatures are also highly correlated with ENSO (Basher and Thompson, 1996; Rhodes et al., 1993) and sea level (Laing et al., 1998). During the El Niño extreme, sea level falls, due mainly to lower water temperatures, and the opposite occurs during La Niña because of raised water temperatures.

ENSO affects the frequency and tracks of extratropical cyclones affecting New Zealand (Basher and Zheng, 1995; Sturman and Tapper, 1996). These are generated as tropical cyclones north of New Zealand, and their formation is more frequent during La Niña conditions. Further, they are more likely to follow paths that affect New Zealand. There also appears to be a different frequency of storms of other origin, and an adjustment in the atmospheric pressure pattern over New Zealand (Gordon, 1985). These changes affect the frequency and magnitude of storm surges (de Lange and Gibb, 2000b).

ENSO influences the distribution of precipitation over New Zealand (Gordon, 1985), with the north-eastern coast experiencing higher precipitation during La Niña extremes than during El Niño extremes. Finally, ENSO affects the distribution of winds over New Zealand (Gordon, 1985; 1986). The El Niño extreme is associated with a northwards shift of the Westerly Wind Belt, increasing the incidence of south-westerly winds. During the La Niña extreme, the subtropical anticyclonic belt moves southwards, increasing the incidence of northerly quarter winds (Sturman and Tapper, 1996).

Table 5. Summary of the observed ENSO extreme effects on the northeast coast of New Zealand.

	El Niño	**La Niña**
Air temperature	Decreased	Increased
Atmospheric pressure	SE to NW pressure gradient	NW to SE pressure gradient
Wind direction	More southwesterly winds (offshore)	More northwest-northeasterly winds (onshore)
Storm frequency	Reduced extratropical cyclone activity	More extratropical cyclone activity
Sea surface temperature	Decreased	Increased
Sea level	Drops	Rises
Wave climate	Reduced sea component	Increased sea component
Wave steepness	Reduced	Increased
Near bed flow	More onshore	More offshore
Coastal response	Tendency to accrete	Tendency to erode

The effect of these changes is significant on the northeast coast. This region of coast has few large rivers providing sediment, an indented coastline, and a relatively low energy wave climate with little seasonal variation. Therefore, much of the sandy coast occurs as pocket beaches, and the larger sandy beach systems are associated with low rates of littoral drift (Harris, 1985; Williams, 1985). Hence, the beach systems are sensitive to small changes in wave climate, nearshore current regime, and sea level. Along the Coromandel Peninsula, La Niña extremes are associated with more intense rainfall events and associated mass movement. Therefore, ENSO may also affect the supply of sediment to the coast.

Very few wave data are available for the New Zealand coast, and most are short duration records collected for specific projects (Hume et al., 1997). However, the existence of inter-annual changes in wave climate have been recognised and attributed to ENSO. These variations have also been linked to adjustments in the morphology of tidal inlets through variations in littoral drift (Hicks et al., 1999). On the north east coast, it is also suggested that larger and steeper wave conditions are more frequent during the La Niña extreme as the overall patter of behaviour appears to be the opposite to that experienced on the west coast of North America.

Combining the observed impacts of recent ENSO extremes, a conceptual model of sandy coast response to ENSO can be constructed for the northeast coast (Table 5). Ignoring the effects of littoral drift and sediment availability, this model assumes that adjustments in shoreline position are driven largely by onshore-offshore sediment transport. Coastal erosion is more prevalent during La Niña extremes due to elevated sea levels, onshore winds and larger, steeper waves.

Analysis of beach profile data for the Bay of Plenty indicates that the beaches do respond in the manner indicated by Table 5 in response to ENSO (Smith and Benson, 2000). However, the same data also show that longer period oscillations are possibly present. Unfortunately the available data do not cover sufficient time to define the character of longer scale oscillations. Instead, it is necessary to consider the longer period fluctuations in the forcing processes identified in Table 3.

Interdecadal fluctuations in the Pacific

In the 1990s, low frequency (12-70 year) climate variability was recognised as an important contributor to observed weather patterns around the globe. In the Pacific Ocean four patterns of low-frequency climatic fluctuations have been identified (Tourre et al., 2001):

- Pacific (inter) Decadal Oscillation (PDO) – variability in climate and ecosystems first identified as affecting Alaska and western Canada (Mantua *et al*., 1997).
- Inter-decadal Pacific Oscillation (IPO) – variability in climate identified for the Indian Ocean, Australia, and the Southwest Pacific Ocean (Power *et al*., 1999).
- Decadal and Interdecadal Climatic Event (DICE) – variability in climate identified for the North Pacific Ocean through variations in the location and strength of the Kuroshio Current and ocean fronts (Nakamura *et a*l., 1997).
- Bi-Decadal Oscillation (BDO) – variability in incidence and severity and drought over the western United States (Cook *et al*., 1997).

All of these oscillations are highly correlated, and probably represent a global oscillation (Tourre et al., 2001). Because IPO is the term applied to interdecadal fluctuations in the Southwest Pacific (Power *et al*., 1999), it is now used to represent all the phenomena listed above.

The IPO has been characterised recently as a sequence of climatic regime shifts associated with interacting bidecadal and pentadecadal oscillations (Minobe, 1997; 1999). Hence, the IPO is the consequence of interacting oscillations, rather than a single oscillation. Due to the short records analysed, there may be longer period (century scale) oscillations involved as suggested independently by Fairbridge (1998). In particular alternate cycles appear to weakly and strongly interact with ENSO.

Minobe (1999) identified climatic regime shifts around 1923, 1948 and 1976. Each climatic regime shift has been observed throughout the northern and tropical Pacific Ocean (Mantua et al., 1997; Minobe, 1997; Zhang et al., 1997). Minobe (1999), by extrapolating the cyclic behaviour determined by wavelet analysis, predicted that the next climatic regime shift should occur between 2000 and 2007. However, the onset of weaker ENSO cycling after 1998 represents another climatic regime shift, starting slightly earlier than predicted.

Decadal changes in rainfall and temperature distributions are evident in New Zealand (Salinger, 1980a; 1980b; 2001; Salinger and Mullan, 1999; Tomlinson, 1992), with three periods being recognised: before about 1950; 1950 to 1976; and 1976 to the 1998. A sequence of cycles extending back to the 1650s is evident in tree ring data around New Zealand (D'Arrigo et al., 1995; Jane, 1983). The data match cycles obtained for tree rings in Alaska that have also been linked to the Aleutian Low and the IPO (D'Arrigo et al., 1999; Wiles et al., 1998).

The IPO appears to modulate the behaviour of Monsoon and ENSO (Torrence and Webster, 1999). Numerical modelling and observational data for the California-Oregon coast suggest that negative IPO phases reinforce La Niña extremes, as well as positive IPO phases reinforcing El Niño extremes (Gershunov and Barnett, 1998). However, the Aleutian Low more directly influences the California-Oregon coast, so the effect in New Zealand may not completely follow the North Pacific pattern.

A review of instrumental data for the last two IPO cycles (negative phase from 1946-1977 and positive phase from 1978-1998) shows that the IPO contributes significantly to variations in sea-level pressure, temperature, and precipitation over the Southwest Pacific (Salinger, 2001). The data also indicate that the positive phase had a stronger affect on atmospheric processes than the preceding negative phase. This is consistent with the analysis of sea level undertaken by Goring and Bell (1999), which indicated a stronger sea level response to ENSO forcing after 1976. Due to the lack of long-term data, it is unclear whether the IPO only has a dominant effect in the Southwest Pacific during positive phases, or if there is a longer-period fluctuation superimposed on the IPO that modulates the IPO influence.

The west coast of North America has better observational data for coastal processes than available in New Zealand. An analysis of the wave climate indicates that it has been strongly influenced by an increased incidence of El Niño extremes since 1980 (Seymour, 1998). This is reflected in a measured increase in ocean wave heights for the Eastern North Pacific since 1976 (Allan and Komar, 2000a). There are also changes in peak wave period and possibly wave direction although wave direction was not explicitly considered. However, it is clear that the increase is not solely due to the increased incidence of El Niño extremes (Allan and Komar, 2000b). It appears that changes in atmospheric circulation, and storm magnitude and frequency associated with the IPO cycle may be responsible, although the data do not span a complete IPO cycle.

Given the similarities (albeit with a 180° phase shift) between processes on the northeast coast of New Zealand and those on the west coast of the USA, it is possible that a similar response to IPO may be expected here. That would indicate an increase in ocean wave heights during a negative phase of IPO. If this is the case, it is likely that AEP values for coastal hazards will vary over the IPO cycle.

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