Geological Society of New Zealand

Annual Conference 2001

27th - 29th November, Hamilton

"Advances in Geosciences"

The University of Waikato
Te Whare Wānanga o Waikato

Earth Sciences
The University of Waikato
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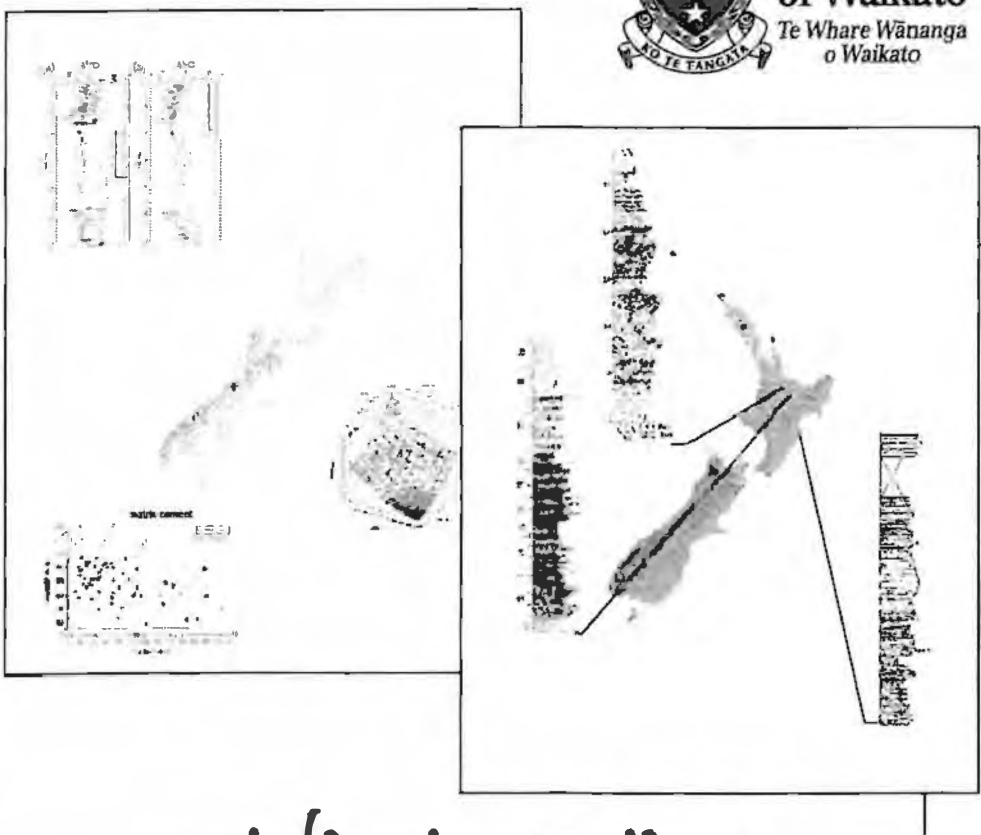

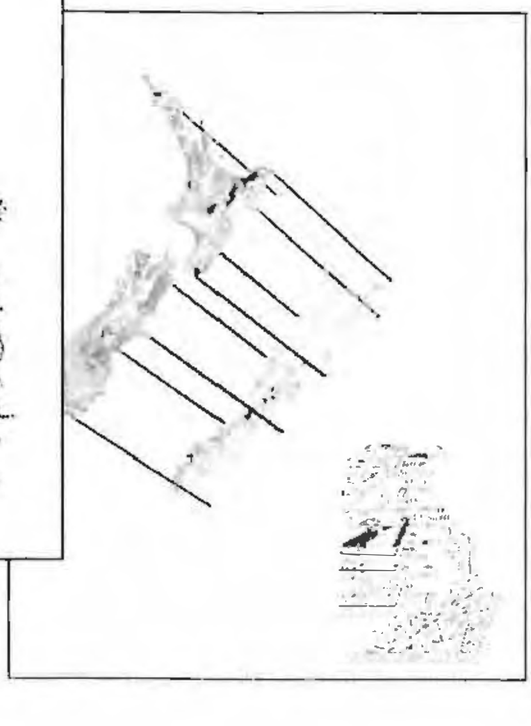

Fieldtrip Guides

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Earth Sciences
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New Zealand

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2001 CONFERENCE FIELD TRIPS

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Acknowledgements

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Field Trip FT3

Coastal Hazards of the Bay of Plenty Coast

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INTRODUCTION

The latest GSNZ fieldtrip to examine coastal processes in the Bay of Plenty ran in 1988. In the intervening 13 years there have been changes in both the legislative framework and our scientific understanding that affect coastal management at the start of the 21st Century. The most important legislative change has been the Resource Management Act (RMA) introduced in October 1991, and the associated Coastal Policy Statement. Amongst other requirements has been the need to define Coastal Hazard Zones for the entire 10-15,000 km of the New Zealand coast.

This field trip (Environment Court permitting) will particularly examine the advances in our understanding of coastal processes and hazards in the context of coastal development in the western Bay of Plenty between Waihi and Little Waihi.

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);
- sea level rise;
- 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
- the 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
- tsunami.

In some areas, the effects of intense rainfall should also be included, particularly where coastal development has modified the catchment and drainage characteristics.

DEVELOPMENTS IN UNDERSTANDING DURING 1980S AND 1990S

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 2001 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; and
- climatic variability and its' influence on the coast

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.

Three surf similarity parameters have been shown to be 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):

the nearshore or breaking Iribarren number ξ_b, which may be expressed as (Battjes, 1975)

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

the surf scaling factor ε_b (Gourlay, 1992; Van Dorn, 1978), defined by

$$\varepsilon_b = \frac{\pi H_b}{L m^2} = \frac{2\pi^2 H_b}{g T^2 m^2}$$

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

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

The surf scaling parameter may be expressed in terms of the Iribarren number, as follows

$$\varepsilon = \frac{\pi}{\xi_b^2}$$

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 (Figure 1).

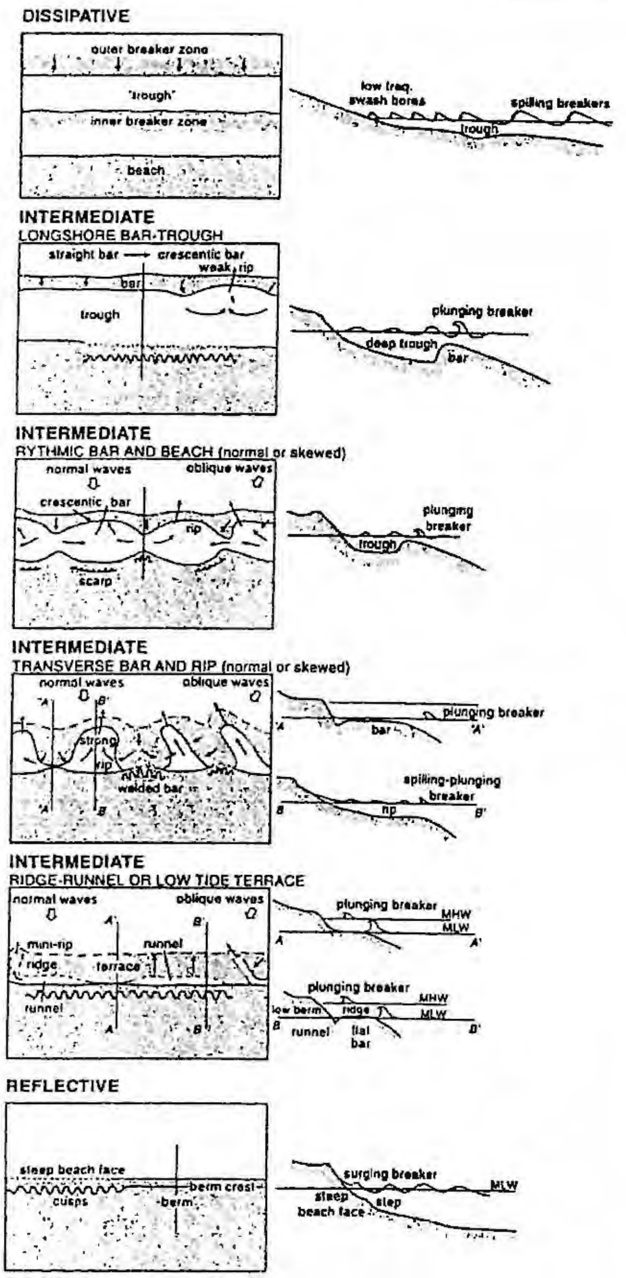

Figure 1. Classification of beach states according to Wright and Short (1983). This model has been applied to New Zealand beaches.

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 (Figure 1), 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. 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 fur-

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

ther 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 (Figure 2) 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.

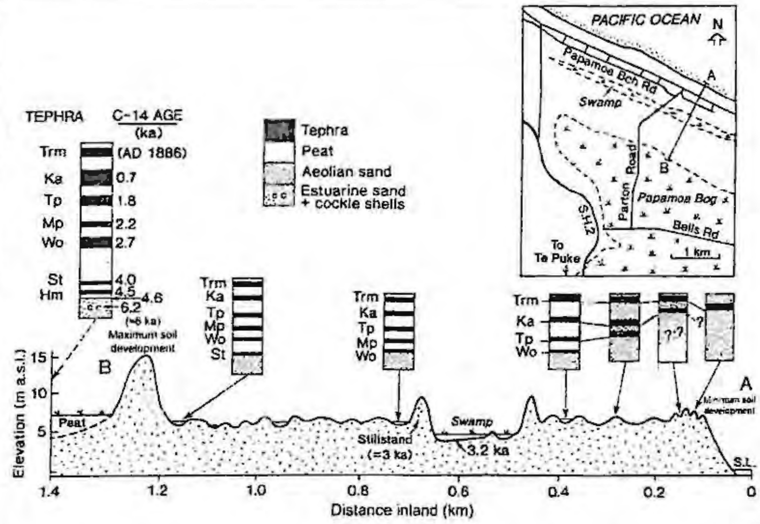

Figure 2. Papamoa beach dunes transect and associated tephrostratigraphy. Tephra abbreviations are: Trm, Rotomahana Mud (Tarawera Tephra); Ka, Kaharoa Tephra; Tp, Taupo Tephra (Unit Y); Mp, Mapara Tephra (Unit X); Wo, Whakaipo Tephra (Unit V); St, Stent Tephra (Unit Q); Hm, Hinemaiaia Tephra (Unit K). After Lowe *et al* (1992) and Alloway *et al* (1994).

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.

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 et al., in prep; 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).

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 (200) and Hogg *et al* (in press). 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

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. 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.

So for example, in 2002 the Moon is at it's closest approach for the year on 27 February, which also happens to coincide with a full moon. As this is also late summer when sea temperatures are high (causing a seasonal rise in sea level), a storm surge around this time can be expected to produce extreme and possibly very damaging water levels. A similar junction of full moon and perigee occurs on 28 March, though the Moon will be 100 km or so further away.

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:

1. Observed storm tide levels around the Bay of Plenty following cyclones Fergus and Drena differed significantly from those measured at Moturiki Island (Blackwood, 1997);
2. 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, 2000).

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, 2000; 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 (2000), 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

Sea level has been rising globally 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.

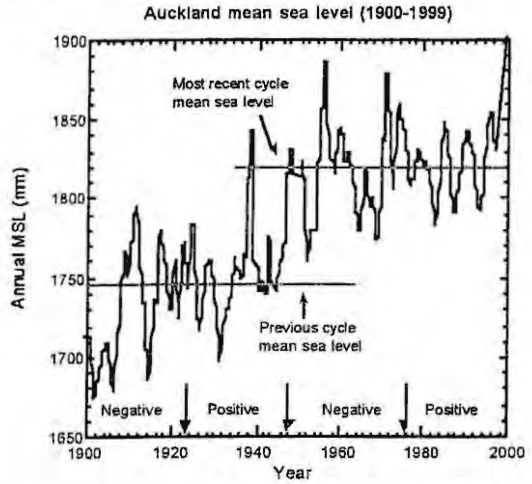

Figure 3. Annual mean sea level recorded at Queens Wharf, Waitemata Harbour, Auckland (after Bell *et al.*, 2000). Also indicated are the mean sea levels for the two 20th-century IPO cycles, and the approximate timing of transitions between IPO phases (arrows). Sea level rose sharply over 1998-99 in response to La Niña conditions. It is presently unclear whether this represents the start of another IPO cycle, or normal interannual variability.

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. The best available long-term record is from Auckland (Figure 3). This record shows that New Zealand sea level rise of 1.7 mm.y^{-1} is consistent with the global average of 1.0-2.5 mm.y^{-1}. More significantly, the record shows that the inter-annual variation is up to half the observed centennial-scale rise, and comparable to lowest predicted sea level rise by 2100 (0.08-0.88 m). At ENSO time-scales (2-7 years), the total variation is even more significant.

There are also decadal-scale variations in sea level evident (Figure 3), suggesting a step-like change in sea level at time-scales of 40-60 years. The possible driving mechanism for this is discussed below. 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.

One aspect of sea level rise that has not made progress over the last 13 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 (and fodder for the lawyers). 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 (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.

However, it is argued that future sea level rise could be larger and faster than the historic changes due to accelerated global warming. The 1987 Edgecumbe Earthquake has provided a useful demonstration of the effect of a rapid increase in sea level (Ruscoe, 1988). Coseismic deformation resulted in a relative sea level rise of ~1 m at the coast, fortunately affecting sites monitored for coastal erosion. Subsequent to the earthquake there has been no shoreline retreat that can be attributed to the rise in sea level (Pickett et al., 1997). For the affected sites it is argued that factors ignored by the Bruun Rule are considerably more important. This situation may well apply elsewhere in the Bay of Plenty.

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, 2000).

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).

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 3). 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 3 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 recently 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 al., 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, it will be used to represent all the phenomena listed above.

Table 3. 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 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).

Minobe (1999) identified phase reversals in the bidecadal oscillation in 1923/24, 1946/47 and 1976/77, and in the pentadecadal oscillation in 1922/23, 1948/49 and 1975/76. These were associated with 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. The onset of a persistent La Niña event in 1999 has been suggested as marking the start of a climatic regime shift. However, it is not yet clear whether this is a correct interpretation.

Decadal changes in rainfall and temperature distributions are evident in New Zealand (Salinger, 1980a; 1980b; Salinger and Mullan, 1999; Tomlinson, 1992), with three periods being recognised: before about 1950; 1950 to 1976; and 1976 to the present. 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 et al., in press). 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.

FIELDTRIP ROUTE

Many thanks to **Dr Jeremy Gibb** for his assistance in guiding today's field trip.

Background information of the geology of the sites visited can be found in:

Briggs, R.M., Hall, G.J., Harmsworth, G.R., Hollis, A.G., Houghton, B.F., Hughes, G.R., Morgan, M.D., and Whitbread-Edwards, A.R., 1996, Geology of the Tauranga Area, Occasional Report No 22: Hamilton, Department of Earth Sciences, University of Waikato, p. 57 + map.

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Omokoroa

View coastal cliff erosion and reserve – issues with cliff stability, sediment supply and storm surges.

Waikaraka

View expansion of mangrove and retreat of seagrass – issues with changes in sediment texture & nutrient status, due to land use changes and/or climatic variability.

Ocean beach

View changes in development type (more intensive, high rise) – issues with beach state, renourishment, coastal erosion, climatic variability

Tay Street

Proposed artificial surf reef site – issues with alternative protection strategies, public perceptions, recreation economics

Papamoa

View Taylor reserve (reclaimed stream bed) – issues with CHZ and risk, perceptions of coastal hazard, effect of storm surges, sediment supply, and climatic variability

Maketu & Little Waihi

View changes due to sediment supply and river diversion – issues with estuarine remediation, CHZs and sand mining.

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