ERRATA

p.6 line 10 : even with
p.17 line 18 : Thus, these uncertainties indicate that quantitative evolutionary models (whether based on
p.21 line 16 : Sheet 10 Wanganui (Lensen 1959)

p.33 line 5 : in producing
p.34 line 19 : base of a
p.47 line 9 : relative to
p.82 footnote : below which occur marine fossils
p.89 4.5.5.4 line 2 : within the Upper Brunswick

p.140 line 3 : inferred to have occurred during glacial epochs in Taranaki
p.161 line 15 : source of streamflow in the surveyed terrace front valleys
p.196 line 2 : originates as subsurface flow at the valley head

Bibliography

The following references were inadvertently omitted:


N.B. All grid references quoted in the text refer to the one thousand yard grid of NZMS1 maps (see Map).
UPPER QUATERNARY LANDSCAPE EVOLUTION IN SOUTH TARANAKI, NEW ZEALAND

by

BRADLEY JOHN PILLANS

Thesis submitted for the degree of Doctor of Philosophy at the Australian National University

Canberra
February 1981.
DECLARATION

Except where otherwise acknowledged in the text, this thesis represents the original research of the author.

B.J. Pillans

B.J. Pillans
Rapanui Terrace at Ohawe Beach
I am pleased to acknowledge the patient assistance of both my supervisors: Dr J. Chappell (ANU) and Professor M.J. Selby (Waikato NZ), whose willingness to help me at all times over the past six years has been greatly appreciated.

I also thank the ANU for financial support, and Professor B.L.C. Johnson and Mr E.C. Chapman (Geography Dept. ANU) for continual support of departmental facilities and attention to administrative details. The Geography Department, James Cook University of North Queensland, also provided financial and laboratory assistance during 1978.

The following people have all made invaluable contributions to the making of this thesis:

Mr L.B. James, John Curtin School of Medical Research, ANU (amino acid analyses)
Dr V.E. Neall, Soil Science Dept., Massey University, NZ (field assistance and discussion)
Dr M.S. McGlone, Botany Division DSIR, Christchurch, NZ (palynological work and discussion)
Dr B.P. Kohn, Ben Gurion University of Negev, Israel (fission track data)
Dr A. Beu, NZ Geological Survey, Lower Hutt NZ (palaeontological identifications)
Mr R.N. Patel, Botany Division, DSIR, Christchurch, NZ (wood identifications)

In addition I greatly benefited from fruitful discussions with Dr D.L. Dunkerley, Dr A.R. Edwards, Sir Charles Fleming, Dr G. Meehan, Dr J.D.G. Milne, Professor J. Soons, Dr R. Summons and Professor H. Wellman.

I am particularly indebted to Rob and Linda Dymond of Hawera, and Neill Kennedy, Rotorua, NZ for accommodation while on fieldwork, especially when the weather turned nasty or my cars broke down. In addition I gratefully acknowledge the cooperation of the many
landowners in Taranaki, who cheerfully allowed access to their properties even if I didn't find any gold, silver, oil, tin, copper....

Without the draughting expertise of Kevin Cowan, and the patience of Gennesse ("I am not a typist") Winch, the finished product would never have been possible.
ABSTRACT

An extensive flight of Upper Quaternary marine terraces in South Taranaki, North Island, New Zealand, is viewed as a giant natural laboratory in which long term rates of landform development may be quantitatively measured. Amino acid racemisation dating, calibrated to the fission track dated Omahina Tephra (370±50Ka BP), enables fossil wood samples from a variety of sites within the terrace sequence to be dated. Twelve terraces are mapped and described, and are estimated to have been formed by high sea level events at 60, 80, 100, 120-135, 210, 310, 340, 400, 450, 520, 595 and 680Ka BP. Five of these terraces have not previously been recognised in South Taranaki; eight are formally named in this study.

Stratigraphic relationships within volcanic deposits in Western Taranaki, and marine sediments within the South Wanganui Basin are discussed. Correlation with overseas data is attempted; in particular correlations with the oxygen isotope record of Pacific core V28-238 (Shackleton & Opdyke 1973) and with coral terrace sequences in New Guinea (Chappell 1974b) and elsewhere appear to be well supported. The implications of the South Taranaki terraces for New Zealand Quaternary chronology are discussed.

Doming-type uplift is demonstrated to have been responsible for terrace deformation. Mean uplift rates vary between 30 and 70cm/Ka. Shore-parallel tilting amounts to c.7nrad/year while shore-normal tilting amounts to c.27nrad/year.

Geomorphologic observations of a simple suite of fluvial landforms demonstrate the need for a geological basis to understanding their long term development. The evolution of stream thalwegs and valley side slopes is described and quantitatively modelled. The mean rate of ground lowering is calculated to be 6cm/Ka in the 27 surveyed small valleys.
## TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frontispiece</td>
<td>i</td>
</tr>
<tr>
<td>Acknowledgements</td>
<td>ii</td>
</tr>
<tr>
<td>Abstract</td>
<td>iv</td>
</tr>
<tr>
<td>Table of Contents</td>
<td>v</td>
</tr>
<tr>
<td>List of Plates</td>
<td>vi</td>
</tr>
<tr>
<td>List of Figures</td>
<td>vii</td>
</tr>
<tr>
<td>List of Tables</td>
<td>ix</td>
</tr>
</tbody>
</table>

### CHAPTERS

<table>
<thead>
<tr>
<th>Chapter</th>
<th>Topic</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>PREFACE</td>
<td>1</td>
</tr>
<tr>
<td>2</td>
<td>INTRODUCTION TO THE STUDY AREA</td>
<td>19</td>
</tr>
<tr>
<td>3</td>
<td>RACEMISATION DATING : METHOD AND APPLICATION</td>
<td>42</td>
</tr>
<tr>
<td>4</td>
<td>STRATIGRAPHY OF THE MARINE TERRACES</td>
<td>68</td>
</tr>
<tr>
<td>5</td>
<td>INTERPRETATION AND DATING OF THE TERRACE SEQUENCE</td>
<td>114</td>
</tr>
<tr>
<td>6</td>
<td>GEOMORPHOLOGY</td>
<td>147</td>
</tr>
<tr>
<td>7</td>
<td>QUANTITATIVE EVOLUTIONARY MODELS</td>
<td>177</td>
</tr>
<tr>
<td>8</td>
<td>CONCLUSIONS AND DIRECTIONS FOR FUTURE RESEARCH</td>
<td>198</td>
</tr>
</tbody>
</table>

### APPENDICES

<table>
<thead>
<tr>
<th>Appendix</th>
<th>Topic</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Amino acid data</td>
</tr>
<tr>
<td>2</td>
<td>Omahina Tephra localities</td>
</tr>
<tr>
<td>3</td>
<td>Mt Curl Tephra : stratigraphic position</td>
</tr>
<tr>
<td>4</td>
<td>Terrace front valleys : maps and long profiles</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Bibliography</th>
</tr>
</thead>
<tbody>
<tr>
<td>Map 1 inside back cover</td>
</tr>
</tbody>
</table>
LIST OF PLATES

FRONTISPICE : Rapanui Terrace, Ohawe Beach

PLATE 2.1 Mt Egmont
2.2 Mt Egmont, Pouakai and Kaitake volcanic centres
2.3 Dissected landscape inland of terraces
2.4 Summit accordance inland to Mt Ruapehu
2.5 Brunswick and Kaiatea terraces
2.6 Whenuakura River Valley
2.7 Retreating coastal cliffs, Ohawe Beach
2.8 Brunswick Terrace : natural exposure
2.9 Ngarino cliff : surface expression
2.10 Ngarino cliff : stream dissection
2.11 Marine sands, Inaha Terrace
2.12 Marine sands, Kaiatea III Terrace
2.13 Active coastal sand dunes

4.1 Omahina Tephra, reference section
4.2 Omahina Tephra, type section
4.3 Rangitawa Stream, type section
4.4 Siltstone conglomerate, Kaiatea I Terrace
4.5 Terangian Stage : new proposed type section
4.6 Upper Brunswick Terrace
4.7 Ngarino Terrace cover beds
4.8 Rapanui Terrace cover beds
4.9 Rapanui/Inaha cliff : coastal exposure
4.10 Rapanui Terrace : coastal exposure
4.11 Inaha Type Section
4.12 Inaha Terrace, Waingongoro River mouth
4.13 Holocene deposits, Waverley Beach

6.1 Terrace dissection
6.2 Geomorphic expression above and below wave cut surface
6.3 Shore-parallel valley, general view
6.4 Seepage heads, shore-normal valley
6.5 Amphitheatre shaped seepage head
6.6 Seepage heads/spur and groove topography
6.7 Amphitheatre basin head
6.8 Terrace front valley, Kaiatea III terrace
6.9 Terrace front valley, Brunswick Terrace
6.10 Valley floor, terrace front valley
6.11 Debris slide, Brunswick Terrace
6.12 Rotational slumps, Brunswick Terrace
LIST OF FIGURES

FIGURE 2.1 Physiography
2.2 Maps and aerial photographs
2.3 Tectonic framework
2.4 Regional geology
2.5 Marine terrace nomenclature
2.6 Marine terrace development
2.7 Climatic data
2.8 Vegetation
2.9 Lapse rate

FIGURE 3.1 D and L amino acids
3.2 Naturally occurring amino acids
3.3 Hydrolysis/sample chromatogram
3.4 Arrhenius plot: allo/isoleucine epimerisation kinetics
3.5 Contamination
3.6 Selection of equilibrium constant: errors
3.7 Amino acid concentration vs D/L ratio
3.8 Threonine vs D/L ratio
3.9 Serine vs D/L ratio
3.10 Aspartic acid vs D/L ratio
3.11 Dating curve
3.12 Paleotemperatures
3.13 Model errors: summary

FIGURE 4.1 Omahina Tephra
4.2 Kaiatea Terraces
4.3 Brunswick Terraces
4.4 Brunswick type section
4.5 Warrengate Section, Brunswick Terrace
4.6 Ngarino Terrace
4.7 Rapanui Terrace
4.8 Inaha Terrace
4.9 Inaha type section
4.10 Younger deposits

FIGURE 5.1 Elementary deformation patterns
5.2 Shore-parallel deformation
5.3 Idealised relation diagrams
5.4 Overseas sea level data: summary
5.5 Uplift rates/deformation pattern
5.6 Relation diagram, South Taranaki
5.7 Summary correlation, and stratigraphic subdivision

FIGURE 6.1 Terrace data
6.2 Drainage initiation and development
6.3 Typical terrace front valley/geometric framework
6.4 Long profiles
6.5 Power curve fits: examples
6.6 Slope data: summary and examples
6.7 Mean slope vs distance downstream and valley depth
6.8 Slope orientation
FIGURE 7.1 Models of stream downcutting and headward retreat
7.2 Models of stream downcutting and slope development
7.3 Model framework
7.4 East-facing valley side slopes - model fits
7.5 West-facing valley side slopes - model fits
7.6 Stream thalwegs - model fits
# LIST OF TABLES

**TABLE 3.1** Variation of reaction half life with temperature
3.2 Factors influencing racemisation rates
3.3 C14 dates and racemisation data
3.4 Racemisation dates : summary
3.5 Wood identifications

**TABLE 4.1** Existing stratigraphic nomenclature
4.2 Sedimentary environmental diagnostics
4.3 Fission track ages : Omahina Tephra, Rangitawa Pumice, Mt Curl Tephra

**TABLE 5.1** Terrace deformation data
5.2 Strandline heights
5.3 Strandline height ratios
5.4 Rapanui Terrace racemisation dates : summary
5.5 Terrace age estimates : simple models
5.6 Sea level chronology : overseas
5.7 Uplift rates : Ngarino, Brunswick, Upper Brunswick terraces
5.8 Uplift models : post-Rapanui terraces
5.9 Uplift models : pre-Kalatea III terraces
5.10 Terrace age estimates : summary
5.11 Wanganui Basin correlations
5.12 Western Taranaki volcanic stratigraphy
5.13 Existing New Zealand Quaternary chronology
5.14 New Zealand Quaternary chronology : suggested modifications

**TABLE 6.1** Morphometric data : terrace front valleys
6.2 Correlation matrix : terrace front valleys
6.3 Log-log multiple regressions : thalweg slice area
6.4 Log-log multiple regressions : normalised thalweg slice area
6.5 Power curve fits, valley long profiles
6.6 Slope data : summary
6.7 Mann-Whitney U-test of slope asymmetry
In 1964, Richard Chorley stated the Huttonian view in geomorphology: "it is after all the most fundamental canon of geomorphology that past landscapes can be most completely understood with reference to present ones", and then continued, rather ambiguously, to state that the corollary was also valid, ie. "From another point of view, it may be argued that present landscapes can be better understood with reference to past ones" (Chorley 1964, p.62).

These strong, but somewhat circular relationships between past and present landscapes provide the raison d'etre for this thesis; a thesis which is largely concerned with quantification and modelling of long term changes in a New Zealand landscape. The conceptual framework, implied by these very general statements, can be more explicitly expressed in terms of three fundamental postulates:

(1) That present landforms have evolved in a mechanistic way, through the continued action of processes which affect the vertical position of every point on the landsurface.

(2) That although surface-affecting processes vary in magnitude and frequency of occurrence, and may be discrete rather than continuous in space and time, the effect of a given process spectrum remains constant through time.

(3) That long-term averages of processes, if measured and correctly related to climatic and tectonic boundary conditions, can be used to construct landform evolution models, which describe the past and explain the present.
The classical Huttonian view, derived essentially from Playfair's (1802) famous observation that valleys have been formed by the streams which lie in them, is implied by statement (1). The uniformitarian principles of Hutton (1795) and Lyell (1830) are restated in (2) in such a way as to emphasise the reality of short-term process variability (cf. Wolman & Miller 1960; magnitude-frequency concepts), which if assessed properly in statistical terms, enlarges upon but does not disqualify the uniformitarian conceptions.

The first two points above, effectively represent long-standing principles rephrased so as to point towards quantitative models (an important, relatively recent development in evolutionary geomorphology), rather than the qualitative hypotheses, concerning landform evolution, upon which the science of geomorphology was founded and initially developed. This matter, in conjunction with statement (3), is discussed further below.

1.1 THE BASIS FOR QUANTITATIVE EVOLUTIONARY MODELS AND PROBLEMS OF THEIR TESTING

The famous qualitative conceptions of landform evolution, stemming from the earliest ideas of Playfair, through Powell, Gilbert, and in particular, the "cycle of landform evolution" theories of W.M. Davis, are deductive: landscapes of various degrees of dissection, relief and steepness were observed in different regions, and were conceptually organised into an evolutionary procession of stages (most clearly articulated in Davis's "normal cycle" and "normal cycle in arid landscapes"). Later authors (eg. Von Engeln 1942) embellished the theory, by similar deductive reasoning, and Davisian views and intellectual method were most highly detailed (to include effects of tectonic and other interruptions of the "normal" evolutionary course), by Cotton in particular (eg. Cotton 1942).
Quantitative argument concerning landform evolution rests on a different intellectual strategy, i.e., the inductive method, where the real or supposed actions of surficial processes are applied mechanistically to an initial or intermediate landscape form, and then are advanced forwards (or sometimes backwards) in time, to predict, through successively-estimated increments, the course of landform evolution. Probably the first accessible author explicitly to use this method was W. Penck, whose graphical methods and results, particularly pertaining to parallel slope retreat and footslope or pediment development, are justly famous in the discipline. Although Penck's general conclusions are faulty, essentially on two grounds:

1. His estimations of processes on slopes certainly do not apply in a wide range of climatic contexts which he discussed (Penck 1953),

2. The graphical method is inherently imperfect owing to the mathematical non-linearity of process-form equations (cf. Scheidegger 1970),

his inductive technique stands as the one major basis upon which the classical models of landform evolution may be investigated.

The quantitative slope development models of Bakker & Le Heux (1952), Scheidegger (1961), Young (1963), Hirano (1968, 1975), Carson & Kirby (1972), Ahnert (1973, 1976), Chappell (1974) and Mizutani (1974) are good examples of the recent trend towards the type of quantitative evolutionary modelling originally pursued by Penck. These models are representative examples of a major class of quantitative evolutionary models - the deterministic models, for which a unique solution is obtained from a specified set of initial conditions, boundary conditions and geomorphic process spectrum acting over a given interval of time. The other major group of quantitative evolutionary models are the stochastic models, which attempt to allow for the certain amount of randomness apparently present in nature. This group, for example, includes the thermodynamic models of Chorley (1962), Langbein & Leopold (1964, 1966), and Scheidegger (1964), as
well as the simulation models of Scheidegger & Langbein (1966), Howard (1971), King & McCullagh (1971), Thornes (1971) and Dunkerley (1977b).

The applicability of the mathematical inductive strategy rests not only on some knowledge of the relationships of process to climatic boundary conditions (cf. "morphogenetic regions" concept of Peltier (1950)), but also on the nature of the past history of these boundary conditions: for example classical geomorphologists have long since shown the importance of Pleistocene climatic changes and of orogenic tectonism on larger-scale fluvial landscapes (eg. in the New Zealand context, Cotton 1942, 1950, 1958a,b). However, note that where a process magnitude-frequency spectrum has highest magnitude-lowest frequency events on similar time scales to Pleistocene boundary condition changes, the inductive problem is intractable.

The viability of modern quantitative evolutionary models rests heavily on two major factors:

1. The testing of the model against natural landscapes.
2. The mathematical representation of the process(es) which is (are) regarded as being significant in a given case. These are now discussed.

1.2 TESTING THE MODEL AGAINST NATURAL LANDSCAPES

1.2.1 INITIAL FORMS

The argument from process to form in quantitative evolutionary modelling requires that certain initial and boundary conditions be specified before a unique solution (form) can be obtained from the mathematical representation of a given geomorphic process spectrum.

Ideally then, as a partial requirement to properly test any model against natural landforms, one looks for a landscape in which are preserved features that allow their initial form to be reconstructed and dated. Comparison of initial and present forms can then
provide us with a measure of average rates and directions of change. Additionally, further constraints can be placed on our model, if we were able to reconstruct what an initial form was like at several known points during its development before the present. Alternatively, we could reconstruct the initial forms of a group of similar landforms of different known ages, and assume they represent an evolutionary sequence.

Unfortunately, very few landscapes preserve within them sufficient evidence with which to make these reconstructions because -

1. dating techniques are materially limited
2. relatively few datable materials are preserved intact in the geomorphologic environment
3. the course of landscape development tends to destroy the record of its previous history.

These problems, when coupled with the slow rates of change in most natural landscapes, therefore represent major barriers to the field testing of quantitative evolutionary models. Consequently, this has led to the relatively popular strategy of extrapolating data from laboratory simulations (eg. Mosley 1973, Parker 1976), or studies of rapidly evolving landforms (eg. Schumm 1956), in order to model the long term development of slowly evolving natural landforms. However, as pointed out by Shreve (1979) there are dangers involved in the averaging of short time scale processes and their extrapolation to longer time spans, owing to the nonlinearity of most geomorphic processes.

Yet another strategy, to circumvent these problems, has been to argue that given a long enough period of time, the influence of initial form may be reduced or even eliminated. (In much the same way, pedologists have argued that the influence of parent material is most strongly manifest in the initial stages of pedogenesis eg. Chesworth 1973). Although this premise of "convergent evolution"
has yet to be adequately tested, it finds expression in the models of Kirkby (1971, 1976a) in relation to the attainment of equilibrium slopes or "characteristic forms", in which the model slopes evolve relatively rapidly to a state that depends entirely on process, and is virtually independent of initial form. Despite the attractiveness of this and other equilibrium models (eg. Hack 1960, Leopold & Langbein 1962, Langbein & Leopold 1964), the fact remains that the open system nature of most natural landscapes makes the attainment of approach to equilibrium virtually impossible (eg. Bull 1975, Shreve 1979). In fact, as Chappell & Eliot (1979) have showed, even the inshore morphology of medium to high energy sandy surf beaches, where morphologic response to changing environmental factors is relatively rapid, disequilibrium is the rule rather than the exception, between morphology and circulation.

Fortunately, in certain favourable landscapes, initial forms may be reconstructed and dated, and quantitative data obtained on their rates and directions of geomorphic change. For example Ruxton & McDougall (1967), Ollier & Brown (1971) and McDougall et al. (1975) all reconstructed relatively simple, dated volcanic landforms to derive quantitative rates of erosion. Extending this approach, Mizutani (1974) tested theoretical models of stream and slope development against relatively undissected volcanoes of various, but unknown ages. Although Mizutani lacked an absolute chronology, he was able to reconstruct initial forms, and was able to show by curve fitting procedures that his mathematical models were statistically good representations of the evolution of natural landforms. Similarly, dated coral reef terraces on the Huon Peninsula of New Guinea, whose initial forms could be reconstructed, provided Dunkerley (1977a, b) and Chappell (1974a) with appropriate field data against which to test their models of stream network and thalweg evolution respectively.
Other initial surfaces which have provided insights into the long term development of certain landscapes include river terraces (Carter and Chorley 1961), glacial till sheets (McConnell 1966, Thomas & Tuttle 1967) and glacial valleys (Thornes 1971).

1.2.2. TESTING OF THE GOODNESS OF FIT

Even assuming that a sequence of simple, dated initial forms exists, there still remains the problem of testing how closely a model describes the evolution of these initial forms. Sadly this aspect of quantitative modelling has largely depended on visual comparisons (ie. the model "appears" to match field examples), if indeed any comparison with natural landforms is attempted. Thus for example, Smith & Bretherton (1972) simply stated that their model "accords with geomorphological observation". Similarly Parsons (1976) claimed that his model "agrees reasonably well" with field evidence, while King & McCullagh (1971, p.37) stated that their simulated spit was "...remarkably similar to the form of the real spit". In other words, qualitative assessments have been used to test the goodness of fit of quantitative models!

In more rigorous comparisons, however, Mizutani (1974) and Chappell (1974a) used statistical curve fitting techniques to assess the goodness of fit of their models of valley incision into simple initial surfaces. Similarly Dunkerley (1977a) compared the metric properties of his simulated stream networks with those of field examples using standard statistical methods, and concluded that his model was an adequate description of field data.
In an earlier study, Ahnert (1970) had attempted to compare model slopes with natural slopes that apparently satisfied his model boundary conditions (homogeneous bedrock, no stream incision at the base of the slope, stable climate), and showed a statistically close correlation between slope profile and soil thickness properties of both the model and field slopes. Unfortunately, unlike Dunkerley (1977a, b) and Chappell (1974a), Ahnert was unable to reconstruct or date the initial forms of his field slopes, and his comparison was therefore of somewhat doubtful validity, unless his field slopes had reached some sort of equilibrium condition. Interestingly, however, Ahnert pointed out that the prediction of both profile form and soil thickness by his model, offered some measure of confidence in the agreement between model and natural slopes. Similarly, Dunkerley (1977b) has argued that because several properties of natural stream networks were adequately described by his model, he could be more certain of its validity.

As pointed out by Chappell (1974a, 1978), however, good statistical agreement between a model and natural landforms does not necessarily validate a model. Rather, it simply serves to indicate a possible process-response relationship, which may be one of several equally well fitting models (although certain models can apparently be eliminated on the basis of their poor agreement with field examples). This situation arises not only because of the problem of equifinality, but also because one of the major weaknesses of deterministic evolutionary models is their failure to predict sampling variance (Kirkby 1976a). Thus, according to Shreve (1975), because natural landforms may be "unique", at best we can statistically test that the model represents the mean of a landform assemblage, or in other words, its most probable state (cf. the use of "average forms" by Mizutani (1974) in the testing of his models against field
examples). Similarly, Shreve (1979) has argued that while certain geomorphic processes could be perfectly deterministic, owing to instability in natural landscapes (which may arise from amplification of random perturbations eg. cutting of rills and gullies), "the models representing the system would still have to be stochastic, because of the impossibility of specifying the inputs with sufficient accuracy" (Shreve 1979, p.168). Thus, one of the conditions of deterministic modelling must be that the system to be modelled be stable to small disturbances (Kirkby 1976a).

1.3. MATHEMATICAL REPRESENTATION OF GEOMORPHIC PROCESSES

In quantitative evolutionary modelling, the action of geomorphic processes may be represented mathematically using -

(i) empirical data,
(ii) theoretical relationships,

or as is often the case, a combination of these two.

Essentially, however, whichever approach is adopted, there are two factors which ultimately determine the goodness of fit of any mathematical model to natural landforms. These are firstly, the degree of simplification incorporated in the model, which is also related to our ability to identify all relevant factors, and secondly, the amount of geomorphic randomness (ie. sample variance or geomorphic noise) present in natural landscapes. Together, these comprise the total unexplained variance which an empirical or theoretical model must seek to minimize.

That these two components of total unexplained variance are difficult to separate, is largely the result of our lack of understanding of the interdependence and complexity of process-response
interactions. Thus to use the example of the hydraulic geometry of streams (Leopold & Maddock 1953, Schumm 1960). Functional relationships which describe the interactions between channel shape, stream velocity, discharge and sediment load, have been derived from numerous field measurements; however, it remains unclear just how much scatter in the data is the result of oversimplification of a complex system (ie. failure to identify and incorporate the effect of certain factors), and how much is the result of so called geomorphic randomness inherent in natural landscapes. What is clear from this discussion, however, is that we are unable to specify the level of determinism that can be expected in a perfect model. Instead we must assume that it corresponds to a state of minimum total unexplained variance (cf. Langbein & Leopold 1966), despite the fact that the uniqueness of this minimum remains to be tested (Chappell 1974a).

In addition to the problems of complexity and geomorphic randomness in natural landscapes, we have already noted the non-linearity of these systems, particularly in relation to amplification of random perturbations. This nonlinearity also produces two further problems -

(1) Nearly all the most powerful mathematical techniques apply to linear systems, and the solution of non-linear models is extremely difficult, unless they can be reduced to linear form.

(2) The combined effect of a set of geomorphic processes (or their mathematical representation)..."is not equal to the sum of the individual effects of the same processes acting separately" (Shreve 1979, p.167). Thus even if we were able to separately model the effect of several geomorphic processes, their combined effect may not be simply additive as Chappell (1974a) and Ahnert (1973, 1976) have assumed in their models.

Once again, it is hardly surprising in the light of the above discussion, that geomorphologists have sought refuge in stochastic models. In this respect particularly, the simplifying assumptions of equilibrium and entropy maximisation borrowed from thermodynamics have appeared most attractive, although it has been recently argued that
there is no sound basis for these approaches in geomorphology, owing to the open system nature of natural landscapes (Davy & Davies 1979). Despite this objection, the fact remains that some general principles are responsible for the apparent equilibrium forms in nature, since particular geometrical properties remain remarkably similar across a wide range of landscapes (cf. for example the hydraulic geometry of streams).

1.4 Long term rates of operation of geomorphic processes

From the foregoing discussion it should be apparent that the construction and testing of mathematical representations of geomorphic processes is the first step in testing the goodness of fit of quantitative evolutionary models, which in turn I have argued, requires a knowledge of past events and processes. However, if it were possible to rely on modern quantitative process measurements to construct and test the mathematical representation of geomorphic processes, and then extrapolate these to long term landscape changes, this "historical" requirement would become redundant. Therefore, in this section and the next, I address some of the problems associated with the measurement and extrapolation of modern geomorphic process rates to infer long term rates and their resultant evolutionary forms.

The importance of a relevant time scale in relation to the interpretation of geomorphic processes and their resultant forms was clearly recognised by Schumm & Lichty (1965) when they distinguished three major spans of geomorphic time: cyclic, graded and steady, corresponding to time periods of the order of $10^6$, $10^2$ and $10^{-2}$ years respectively. Using these subdivisions, they were able to highlight two of the major barriers to successful extrapolation from short term data. The first of these, was that the factors which determine the character of landforms can either be viewed as dependent or independent variables as the temporal and spatial dimensions
change: "...as the dimensions of time and space change, cause-effect relationships may be obscured or even reversed, and the system itself may be described differently" (Schumm & Lichty 1965, p.110). Thus, for example, during steady time Schumm and Lichty considered hillslope morphology to be independent of the processes acting upon it, and only during longer time spans (graded and cyclic) did they envisage a dependence of form on process. Similarly, they argued that stream flow characteristics during steady time were dependent on channel morphology, and not vice versa. Therefore, helicoidal flow, for example, was seen to be conditional upon the prior existence of meanders, rather than their cause as has sometimes been suggested (eg. Tanner 1960, Sakalowski 1974). Clearly then, any attempt to model cyclic or graded time process-response relationships based on steady time observations, is unlikely to take account of all causal links and feedback mechanisms.

The second flaw outlined by Schumm & Lichty (1965) was that, while a system may appear to be in equilibrium for short periods of time (cf. the definition of "steady time"), it may still be part of an undetected long term trend away from that equilibrium. In fact Bull (1975, 1979) has stated that the attainment of equilibrium or steady state for long periods of time may be unlikely, and that the use and identification of thresholds may be a more versatile approach. (A geomorphic threshold is seen as a transition point, or period of time, separating different modes of operation of geomorphic processes). Schumm (1977) has recognised two types of geomorphic threshold:

1. **Extrinsic thresholds**, in which a change in a geomorphic system results from external influences eg. climatic change, human impact, tectonism.

2. **Intrinsic thresholds**, which are the result of landform change through time to a condition of instability, without a change in external influences eg. progressive weathering that eventually results in slope failure, meander cutoff resulting from increasing sinuosity, or some glacial surges,
and clearly any long term assessment of geomorphic rates must take account of these thresholds.

Recognition of geomorphic thresholds does not of course mean that processes and landforms cannot proceed towards equilibrium, and Bull (1975) has stressed this point in relation to allometric change in land forms. Allometry, according to Bull, does not require that steady state be reached or closely approximated, but rather, stresses the "tendency for adjustment between materials, process and form, even where steady state does not exist" (Bull 1975, p.1491). Thus he argued that "allometric adjustment can be shown to exist by demonstrating consistent and statistically significant interrelations between two or more geomorphic features of an open system" (p.1497). Although this definition of allometric adjustment is very general, and quite different from the original definition of "allometric growth" used by biologists (eg. Huxley 1924, Gould 1973), some confusion has arisen between the two (see Mosley & Parker (1972), for earlier criticism of the allometric growth concept in geomorphology). Despite this confusion, Bulls' approach in stressing the interdependent adjustment between variables, even in the absence of a steady state, is appealing because it attempts to provide a general description of functional relationships in geomorphology by arguing that they "...result from orderly changes through time in geomorphic systems" (Bull 1975, p.1498).

Unfortunately, as yet, geomorphologists have very little information pertaining to the rates of adjustment and the nature of the orderly changes to which Bull refers. This is particularly true of those geomorphic thresholds which occur only rarely (eg. once or twice in graded or cyclic time spans) and which may never be experienced in a human lifetime (see magnitude/frequency concepts
Thus, although systems theory tells us (Chorley & Kennedy 1971) that associated with these thresholds, there are characteristic reaction times and relaxation times, which may vary with

1. the magnitude and direction of rate of change,
2. the resistance of the system to change,
3. the complexity of the system,

too few data are at present available to geomorphologists which adequately allow for these variables in extrapolation from short term to long term rates.

To summarise, the principle reason why present process rates (as we measure them) cannot be confidently extrapolated to the long term, is likely to be a result of the shortness of our observations and our inability to obtain a representative sample of long term trends. Estimation of the long term rates of operation of geomorphic processes is therefore better approached through the method of reconstruction and dating of initial landforms (cf. the classic study by Ruxton & McDougall (1967)).

1.5 UNIFORMITARIANISM

As Gould (1965, 1967) and Hubbert (1967) have indicated, uniformitarianism, as we presently understand it, embodies two fundamental postulates:

(1) A substantive uniformitarianism, which argues that past geological events and processes have proceeded at approximately the same rates as those we observe today.

(2) A methodological uniformitarianism, which argues that
(a) natural laws are spatially and temporally invariant, and
(b) that no hypothetical unknown processes need be invoked, if observable historical changes can be explained by presently observable processes.

The postulates of methodological uniformitarianism are, of course, not peculiar to geological science, but rather are common to all sciences, with the later (2b) being conditional upon the former (2a). It is this composite postulate which forms the basis for
modelling past geomorphic changes in terms of the processes we see today, although as Gould (1965, p.227) pointed out, there is no a priori reason to deny that "...other natural processes now inoperative were then effective". In other words, methodological uniformitarianism is an affirmation of simplicity, since it postulates that no unnecessary theoretical processes need be invoked as long as observable ones can successfully explain past changes. Thus in acknowledging that uniformitarian principles apply, I refer only to the principles of methodological uniformitarianism and concede that while the principle of substantive uniformitarianism is not necessarily valid, it is also not a necessary prerequisite for successful quantitative evolutionary modelling.

If the definition of methodological uniformitarianism is rigidly adhered to, however, without regard for scale of observation, it seems clear that observable historical changes will not be adequately explained by presently observable processes in all cases. For example, when consideration is made of climatic changes, it is clear that the principles of methodological uniformitarianism only remain valid on a global scale. The whole basis of climatic geomorphology is encompassed within the idea that climate controls process which in turn controls form (Derbyshire 1976). Thus if an area experiences a change in climate, a geomorphic threshold may be crossed which allows the mode of operation, interaction and relative importance of geomorphic processes to vary accordingly. Further, if the magnitude of the change is large and/or the scale of observation small, processes entirely new to the area under observation might also be initiated. However, as in the case of the wide occurrence of relict permafrost features dating from the last glaciation, no hypothetical unknown processes need be invoked, to explain the existence of these features in areas where permafrost is no longer present, because
(a) we can observe similar active features elsewhere on the earth's surface at the present time; and

(b) we have many independent lines of evidence which support our requirement for substantial climatic changes since the last glaciation.

In 1960, Wolman & Miller produced further support for uniformitarian principles, when they observed that, particularly in the case of fluvial transport, events of moderate size and frequency appeared to account for much of the material being moved over "long" periods of time. They argued that while extreme events of high magnitude may well cause gross amounts of change in the landscape, their rarity meant that, relative to overall rates of modification, their effect was less. More recently, Wolman & Gerson (1978) have modified these ideas and stress that "...both time and magnitude have significance only in terms of the continuous processes typical of a given climatic region" (p.206). Thus, an extreme event of similar magnitude and frequency may have vastly different effects according to the climatic regime under which it occurs (see also Starkel 1976), and relating to the ability of the landscape to recover the form existing prior to the event (Wolman & Gerson 1978). Clearly then, the magnitude/frequency concepts of Wolman & Miller (1960) and Wolman & Gerson (1978) do not invoke new processes, but emphasise the relative importance and effects of events of differing magnitude in the processes we observe today.

Unfortunately, it is also apparent that the short time span over which we have reliable direct human observations of geomorphic processes and events, limits our experience of the effects and recovery periods of really catastrophic, very low frequency events (Derbyshire 1976). Thus it is unlikely that modern measurements of process rates can adequately allow for these events, particularly those extreme meteorological events which create quite different new forms, or accelerate natural rates of change long after the event has
occurred (Starkei 1976). The Madang earthquake of 1970 (Pain & Bowler 1973) is another example of the type of catastrophic event about which we have very little information. Certainly geomorphic rates derived from a few years observation (eg. erosion rates of Douglas (1969)), cannot hope to represent those of the long term, and it is therefore not surprising that discrepancies arise when comparing modern and long term rates (eg. Dunkerley 1977a), even in those rare cases where the effects of man can be discounted.

In summary, it is apparent from the discussion in this, and the preceding section, that we have no basis for assuming that past geomorphic events and processes proceeded at similar rates to those we observe today. Even if this were true, the short duration of modern process observations prevents the extrapolation of their measured rates to much longer time spans. In addition, although the simplistic assumptions of methodological uniformitarianism provide a sound basis for quantitative evolutionary modelling, we cannot necessarily assume that the mode of interaction and operation of modern day processes was similar in the past. (whether based on empirical or theoretical relationships) require adequate testing against natural landforms, and that this is best achieved using the method of reconstruction and dating of initial forms.

1.6 IMPLICATIONS FOR THIS STUDY

A major shortcoming with quantitative evolutionary models in geomorphology is the lack of their adequate testing against natural landscapes. In the light of the preceding discussion, I will attempt to construct and assess some simple, quantitative evolutionary models by testing them against a New Zealand landscape. The extensive flight of Upper Quaternary marine terraces in South Taranaki, North Island,
New Zealand (see Fig. 2.1) which I will describe, represents a sequence of relatively simple, initial landforms whose ages can be determined. This not only enables testing of models by the method of reconstruction and dating of initial forms, but also provides important quantitative data on the long term rates of operation of a variety of surficial processes within a well defined set of initial and boundary conditions.

Extrapolation from modern process measurements will not be attempted, for reasons outlined above; however, some geomorphic process observations are made in order to more closely constrain the choice of models. In all cases, the goodness of fit of the models to the natural landforms will be quantitatively assessed; good agreement between several model and natural attributes will be taken to indicate greater acceptability of the model.
CHAPTER 2

INTRODUCTION TO THE FIELD AREA

2.1 SETTING

Remnants of marine terraces occur along much of the present New Zealand coastline, and have been described by various authors eg. Fleming (1953), Brothers (1954), Suggate (1965), Chappell (1970, 1975), Ghani (1978) and Yoshikawa et al. (1980).

The best developed set of such terraces occurs as a broad flight on South Taranaki-Wanganui coastal plain in the southwest corner of the North Island. Here, several terraces run in a NW-SE direction, roughly parallel to the present coastline for over 100 km, rising steplike to over 300 m above present sea level and extending up to 20 km inland. Southeast of Wanganui (39°56'S, 175°03'E) the terrace sequence has been largely obscured by extensive fluvial deposits from major rivers draining southwards from the centre of the island, particularly the Rangitikei, Turakina, Whangaehu and Wanganui Rivers (Fig. 2.1). Northwest of Hawera (30°36'S, 174°17'E) the Taranaki Peninsula is dominated by the dormant volcanic cone of Mt Egmont (2518m), which is the second highest peak in the North Island (Plate 2.1). North of Egmont lie eroded remnants of two extinct volcanoes, Pouakai (1399m and Kaitake (684 m), both of which were active during the Pleistocene (Plate 2.2). Consequently, volcanic debris (principally ash and lahars) from these volcanoes has buried the terraces to the north and west of Hawera. Thus, the bulk of the study area lies between Hawera and Wanganui and from the coast to about 20 km inland.
Immediately inland of the marine terrace sequence, the country is highly dissected, with narrow ridges, steeply sloping rectilinear valley sides, and deeply entrenched river valleys with little or no floodplain (Plate 2.3). Summit heights gradually rise inland and appear to form a broad surface on which the base of Mt Ruapehu rests (Plate 2.4). Grange (1927) suggested that the general summit accordance was evidence of an uplifted late Tertiary peneplain which truncated older tertiary sediments at a low angle, although this has yet to be substantiated. Mt Ruapehu (2797 m), an active volcano and the highest peak in the North Island rises impressively to some 2000 m above this surface.

Within the study area itself, drainage is approximately normal to the coast, with several major rivers (Waingongoro, Tangahoe, Whenuakura, Manawapou, Patea and Waitotara Rivers) originating to the north and crossing the terrace sequence in incised valleys containing wide floodplains and prominent river terrace remnants (Fig. 2.1, Plate 2.6). Between the major rivers are broad, flat interfluve areas on which are preserved the marine terrace remnants. These broad interfluves are in turn dissected by smaller streams in steep sided (c.30° slopes) V-shaped valleys (Plate 2.5) with the amount of dissection increasing up to terrace flight until it merges with the completely dissected interior region (Plate 2.3). The drainage patterns of these smaller streams are typically dendritic to trellis, and contrast markedly with the centripetal drainage patterns around Mt Egmont and Mt Ruapehu, and to a lesser extent with the feral (fine textured) stream networks (Cotton 1962) developed on Waitotaran (Pliocene) marine sediments immediately inland of the terraces (Plate 2.3).
2.2 MAPS AND AERIAL PHOTOGRAPHS

A variety of topographical, geographical and bathymetric maps provide a complete or partial cover of the field area (see Fig. 2.2), of which the following are the most important.

(a) NZMS 1 Topographical Series (1:63,360)
   Sheet N119 Stratford
   Sheet N129 Hawera
   Sheet N130 Waitotara
   Sheet N137 Waverley
   Sheet N138 Wanganui

(b) NZMS 18 Topographical Series (1:250,000)
   Sheet 7 Taranaki
   Sheet 10 Wanganui

(c) Geological Map of New Zealand (1:250,000)
   Sheet 7 Taranaki (Hay 1967)
   Sheet 10 (Lensen 1959)

(d) N.Z. Geological Survey Bulletin 52 (Fleming 1953)
   Sheet N137 Waverley (1:63,360)
   Sheet N138 Wanganui (1:63,360)

(e) N.Z. Geological Survey Miscellaneous Series Map 5 (1:1,000,000)
   Quaternary Geology - North Island 1973

(f) N.Z. Hydrographic charts (1:200,000)
   Sheet NZ45 Port Taranaki to Patea
   Sheet NZ46 Wellington to Patea

(g) NZMS 3B Aerial Photograph Series (1:69,000 approx.)

2.3 TECTONIC FRAMEWORK

In the early Tertiary (60-80my BP), separation of the New Zealand continental block from Antarctica (Molnar et al. 1975) and simultaneous opening of the Tasman Sea (Weissel & Hayes 1977) resulted in the northward drift of New Zealand relative to Antarctica. Since the death of the mid-Tasman spreading centre 60my BP, however, New Zealand has become part of the Australian lithospheric plate and since at least 20my BP (Ballance 1976, Nelson & Hume 1977), and possibly
since 38 my BP (Molnar et al., 1975; Carter & Norris 1976), has been underthrust from the east by the Pacific Plate. Thus the character of geological development in New Zealand, since at least mid-Tertiary times has been strongly influenced by its close association with the Australian/Pacific plate boundary.

At present time the boundary between the Australian and Pacific lithospheric plates is marked by the Tonga-Kermadec Volcanic Arc to the north of New Zealand (Karig 1970) and by the Macquarie Ridge complex to the south (Johnson & Molnar 1972). Within New Zealand itself, there is a broad zone of deformation which strikes roughly northeast through both islands linking these two features. This zone is defined by the major historical earthquakes, by active faults, by minor seismicity and strongly tilted Tertiary strata (Clark et al. 1965), as well as the main axial mountain ranges of both islands. As Walcott (1978) has emphasised, the pattern of crustal seismic activity in the New Zealand region indicates that the boundary between the Australia and Pacific Plates is a wide zone of distributed deformation rather than a discrete discontinuity in crustal rocks.

The present estimated direction and rates of relative movement at the plate boundary in the New Zealand region are given in Fig. 2.3, and are based on the instantaneous rotation pole (59.8°S, 178°E) for the Pacific/Australia Plate pair determined by Minster et al. (1974). These rates and directions represent the principal axis of horizontal compression and are the same, within the limits of error, as those determined by Walcott (1978) based on geodetic data and earthquake fault plane solutions. Convergence between the two plates in the South Island is presumed to be accommodated by transcurrent movement along the Alpine Fault (which has a total lateral offset of nearly 500 km (Wellman 1973), as well as vertical movements in the Southern Alps. In the North Island, the situation is less well
Although the features associated with the Tonga-Kermadec Volcanic Arc apparently extend into the North Island (Hamilton & Gale 1968, Hatherton 1970) they become less clearly defined and are separated by a seismic gap to the north of New Zealand (Hamilton & Gale 1968). The Tonga-Kermadec Arc is a complex feature (Karig 1970), consisting of a trench (Tonga-Kermadec Trench), a frontal arc (Tonga-Kermadec Ridge), an interarc basin (Lau-Havre Trough) and a rear arc (Lau-Colville Ridge) (see Fig. 2.3). Analysis of the pattern of deep earthquakes in this region, and their focal mechanisms (Isacks et al. 1968, Johnson & Molnar 1972) indicate that the Pacific Plate is descending beneath the Australian Plate in a subduction zone which dips westwards from the Tonga-Kermadec Trench and extends down to a depth of about 600 km (Isacks et al. 1968). Consequently there is good reason to expect that the Pacific Plate also underthrusts to the west in the North Island.

A laterally inhomogeneous mantle velocity model has recently been proposed by Adams & Ware (1977) for plotting the location of deep earthquakes (>100 km) beneath the North Island of New Zealand, and indicates a uniformly 50° westward dipping Benioff Zone in the western half of the island. Previous solutions using a homogeneous velocity model (eg. Hamilton & Gale 1968) indicated a marked steepening of the Benioff Zone with increasing depth. Both models, however, clearly show that the maximum depth of occurrence of these shocks diminishes southwards from about 350 km beneath the Bay of Plenty, to about 150km beneath the northern part of the South Island. The pattern of seismic activity suggests that the westward dipping Benioff Zone extends at least as far south as, and may be truncated by the Alpine Fault System (Hamilton & Gale 1968, Hatherton 1970, Arabasz & Robinson 1976, Adams & Ware 1977). This southwards shallowing of the maximum depth of
seismic activity beneath the North Island is part of a trend which extends from the northern end of the Tonga-Kermadec Arc and might be explained in terms of either

(a) the southward decrease in rates of plate convergence (see Fig. 2.3)

or

(b) progressive southward propagation of the Kermadec subduction zone in its less well defined southern extension, the Hikurangi Trench (Walcott 1978).

Between the southern end of the Hikurangi Trench and the Alpine Fault, the plate boundary is not well defined. According to Arabasz & Robinson (1976) shallow earthquakes (<20 km) with strike slip mechanisms and ENE-tending slip vectors, support the interpretation that the present Australian/Pacific boundary can be traced as a broad, complex zone of transform faulting connecting these two features. Further north, the plate boundary is interpreted to be near the base of the western wall of the Hikurangi Trench (Arabasz & Robinson 1976, Ballance 1976), although the main region of downturning of the Pacific Plate appears to be beneath the central North Island some 250 km to the west (Hatherton 1970, Adams & Ware 1977). Consequently a broad, shallow segment of the Benioff Zone is inferred to be present under the eastern half of the North Island, and is evidenced in this region by the occurrence of subcrustal earthquakes in the depth range 30-100 km, and by their marked absence of greater depths (Adams & Ware 1977, Smith 1979). The complex deformation on the eastern continental slope (Houtz et al. 1967) and the thrusting mechanism of the 1966 Gisborne earthquake (Johnson & Molnar 1972) also support this interpretation. The eastern North Island thus appears to float on the shallow Pacific Plate lithosphere passing beneath it, and has been designated the Hawkes Bay Crustal Microplate (see Fig. 2.3) by Ballance (1976). As Hatherton (1970) observed, the Taupo Volcanic Zone (see Fig. 2.3), also separates a region of predominantly high heat flow and positive gravity anomalies in the west of the island, from a region of predominantly low heat flow and negative gravity anomalies in the east.
Quaternary volcanism in New Zealand has been confined to the west and centre of the North Island, and much of this appears to be related to a westward dipping subduction zone. Throughout the Cainozoic, basaltic and andesitic volcanism occurred along a wide belt extending from Mt Egmont in the south, through Auckland to the northern extremity of the North Island. The Taupo Volcanic Zone (Fig. 2.3) is a younger linear feature superimposed obliquely across the older western belt, and has been active largely during later Quaternary time (Cole & Nathan 1976). Eruptive rocks within the Zone are dominantly rhyolitic, with isolated andesite stratovolcanoes. Rhyolite domes and associated ash-flow deposits dominate the central part of the Zone, and this area was probably the source of ignimbrite sheets which have flowed over the margins and cover a wide area of the central north island (Nathan 1976). The pattern of volcanism has been interpreted by Ballance (1976) as representing an eastwards and southwards migration of a magmatic arc since about 20my BP.

The narrow strip (25-40 km wide) of recently active vents within the Taupo Volcanic Zone lies directly in line with the Southern end of the Kermadec Ridge (Fig. 2.3). Both Ballance (1976) and Ewart et al. (1977) have viewed the Zone as the logical southward extension of the Tonga-Kermadec Ridge, while arguing that the Lau-Havre Trough and Lau-Colville Ridge have no landward correlatives. Gross differences in the chemical composition of eruptives in the Tonga-Kermadec Arc (basaltic andesites and tholeiites) and the Taupo Volcanic Zone (rhyolites and ignimbrites) are explained by Ewart et al. (1977) as arising from partial fusion of continental crust in the Taupo Volcanic Zone and mixing with parent basaltic andesite magmas. On the other hand Karig (1970) and Walcott (1978) have pointed out that the Taupo Volcanic Zone, which is also characterised by intense
normal faulting, is more likely to be the southward extension of the Lau-Havre Trough, since both are zones of active crustal extension. Unfortunately neither of the above correlations sheds any light on the abrupt southern limit of the Taupo Volcanic Zone: all signs of recent volcanic activity disappear just south of Mt Ruapehu, while the intense faulting also appears to die out in the same area. However, since the South Taranaki terraces lie immediately south and to the west of this critical area, it is clear that the patterns and rates of terrace deformation there will place strong constraints on any tectonic model which seeks to project the features of the Taupo Volcanic Zone southwards through eastern Taranaki (e.g. western margin of the Hawkes Bay Crustal Micropolate, Ballance (1976)).

In summary, much of the deformation in the North Island since at least mid-Tertiary times has its origin in the lateral compressive stress of plate convergence. Late Quaternary uplift along the South Taranaki-Wanganui Coast is likely to be related to this lateral stress, although as elsewhere in the North Island, the precise nature of this relationship remains unclear (Chappell 1975).

2.4 REGIONAL GEOLOGY

The general geological setting of the southwest part of the North Island of New Zealand is given in Figure 2.4.

Much of the Taranaki Peninsula is constructed of andesitic volcanic deposits of the Egmont, Pouakai and Kaitake volcanic centres (Plate 2.2). These centres display a linear southerly progression in ages from Kaitake in the north (dated at about 575 ka BP), to Pouakai (last active between 216 and 250 kaBP), to Egmont in the south which
first became active about 50-70 ka BP and last erupted c.1755 AD (Stipp 1968, Neall 1979). These high K calc-alkaline Taranaki andesites contrast with the low K calc-alkaline andesites of the Taupo Volcanic Zone. The differing chemical compositions have been explained by Hatherton (1970) and Ballance (1976) in terms of their differing heights above the Benioff Zone (225 km in Taranaki and 130 km in the Taupo Volcanic Zone), although the isolated occurrence of such behind-arc volcanic activity, when conditions for its generation should have been favourable right along the strike of the Benioff Zone, remains unexplained*. A small number of very deep earthquakes at about 600 km depth in Taranaki are thought to be caused by a detached slab of the subducting plate (Adams & Ferris 1976, Christoffel & Calhaem 1973), since they lie approximately 250km below the lower margin of the seismically continuous Benioff Zone, and are separated from the main zone of earthquakes by a seismically quiet region.

Associated with the Taranaki volcanoes are extensive ring plain deposits of which lahars (mudflows) are a significant component (Grant-Taylor 1964, Hay 1967, Neall 1979). Dark minerals (eg. titanomagnetite, augite and hornblende) derived from the dissection of these volcanoes and their associated ring plains, produce the characteristic black sand beaches (Frontispiece) of Taranaki (Gow 1967). Similar dark minerals are also present in the marine sands within the uplifted marine terraces - see Plates 2.11 and 2.12.

Underlying the Taranaki volcanic chain and extending seawards beyond the edge of the continental shelf is the Taranaki Basin, which is filled with up to 7000 m of Cretaceous-Cainozoic sediments (Katz 1974, McBeath 1977). New Zealand's only commercial hydrocarbon re-

---

* Further north, recent basaltic volcanism near Auckland (eg. Rangitoto - see Fig. 2.3) has been suggested by Ballance (1976) as originating independently of the subducting lithospheric slab, perhaps from a mantle plume or hot spot.
serves (Maui and Kapuni gas-condensate fields - see Figure 2.4) occur in Eocene sandstone reservoirs within the basin (McBeath 1977). The Taranaki Basin is separated from the younger South Wanganui Basin to the east by the horst of the Patea-Tongaporutu High, which is bounded on its western side by the Taranaki Fault (vertical throw up to 7000m) and on its eastern side by the Strathmore Fault (McBeath 1977). The major N-S trending faults in the Taranaki Basin (Fig. 2.4), which are all normal faults, are depicted by Robinson et al. (1976) and McBeath (1977) as extending south to the northern coast of the South Island. Of these faults, only the Cape Egmont Fault displaces recent sediments, while the others do not cut sediments younger than Upper Miocene (Robinson et al. 1976).

To the east of the Patea-Tongaporutu High lie the two Wanganui sub-basins, the sediments of which unconformably underlie the South Taranaki marine terrace sequence. The North Wanganui Basin is the larger and older of the two sub-basins, and contains up to 7500 m of Oligocene to Pliocene marginal marine and shallow marine sediments (Nelson & Hume 1977). The South Wanganui Basin contains up to 4800 m of Plio-Pleistocene shallow marine sediments (Seward 1976). Cope & Reed (1967) indicated that a structural high, which they named the Pipiriki High, might separate the two basins, but interpretation of regional gravity data (Hunt 1980) does not support this.

The sediments of the South Wanganui Basin have in general suffered only mild deformation. Regional dips of 4 to 7° are common, and dip towards Cook Strait where the inferred basin centre coincides with the highest negative gravity anomaly in the New Zealand region (Robertson & Reilly 1958) - see discussion below. According to Lensen (1959), evidence from oil wells and seismic surveys suggests that the basement has a block-faulted structure, and the South
Wanganui Basin may therefore be considered to be a complex graben structure. Several wide, shallow north-south trending foldings are present with the basin, some of which have been active during the Quaternary (Te Punga 1957b—See Fig. 2.4). Te Punga (1957b) and Lensen (1959) argued that these younger anticlines and synclines were drape folds caused by intermittent block-faulting of the basement.

Several small, but prominent NE-SW trending normal faults cross the South Wanganui Basin. Like the gentle folds described above most have been active during the late Quaternary and displace terrace surfaces (both fluvial and marine) along their trace. The largest of these faults is the Nukumaru Fault Zone which has a maximum known throw of 500 feet (150 m) in Pliocene sediments (Fleming 1953). The maximum measured vertical displacement of the marine terraces is about 50 m on the Nukumaru Fault, and decreases systematically on younger terraces indicating continued movement throughout much of the late Quaternary.

Fleming (1953) provided a detailed account of the geological history of the South Wanganui Basin, and indicated that while the total thickness of sediment in the basin steadily increased, shallow marine to marginal marine conditions persisted throughout its history. Isopach diagrams (Fleming 1953 p.293) indicate that the basin shrunk progressively during its history, and also show an eastwards migration of the axis of sedimentation (Lensen 1959). Nelson & Hume (1977) have suggested that anticlockwise lateral bending of the western North Island during the Pliocene (postulated by Ballance (1976)) may have led to the final uplift and death of the North Wanganui Basin and may have been partly compensated for by rapid down warping in the South Wanganui Basin.

* Ballance (1976) has interpreted the 70° angular discordance between the Taupo Volcanic Zone and older linear belts of volcanic activity in the North Island to be the result of anticlockwise lateral bending about 3my BP.
As indicated above, the South Wanganui Basin coincides with the largest negative gravity anomaly in the New Zealand region. This anomaly forms the southern end of a prominent belt of gravity anomalies extending northeast through the North Island from Cook Strait to East Cape (Robertson & Reilly 1958). The belt is displaced to the west, and substantially isolated from a similar belt of negative gravity anomalies associated with the Tonga-Kermadec Trench to the north (Hatherton & Sym 1975). As Robertson & Reilly (1958) pointed out, the Wanganui Basin gravity anomaly is not explainable in terms of the thickness of sediments in the basin, but must be related to a "predominantly deep seated origin". Hatherton (1970) related the belt of negative gravity anomalies to density contrasts within the mantle, and in particular within the subducting lithospheric plate beneath the North Island. However, the belt of negative gravity anomalies lies at an angle to both the Taupo Volcanic Zone (the main region of downdropping of the subducting slab) and the Hikurangi Trench (the surface trace of the plate boundary), suggesting that the feature is not straightforwardly related to the subducting plate.

Recent fission track dating of rhyolitic volcanic ashes which occur within the shallow marine sediments of the basin (Seward 1974, 1976, Boellstorff & Te Punga 1977) indicates that the South Wanganui Basin was receiving volcanic detritus spasmodically throughout much of the Pleistocene and part of the Pliocene (many of these ashes are likely to be distal equivalents of ignimbrites erupted in the Taupo Volcanic Zone). On the basis of these dates, Boellstorff & Te Punga (1977) estimated that marine deposition in the basin finally ceased by about 340 Ka BP, and that the basin has been progressively uplifted since that time. Robertson & Reilly (1958) have indicated that the Wanganui Basin is not in isostatic equilibrium, which suggests that final uplift and death of the basin may have been, at least in part
isostatically controlled. In this respect the relationships between the older marine terraces and the younger marine sediments of the basin are of critical importance. Fleming (1953) first suggested that some of the higher (older) terraces in South Taranaki might be the shoreward equivalent of the youngest Castlecliffian sediments in the centre of the basin. Dating and stratigraphy in this study confirm this suggestion which is discussed fully in a later section.

2.5 GENERAL GEOMORPHOLOGY

The morphology of the present South Taranaki-Wanganui coastline is characterised by high cliffs up to 60 m in height, and extensive wave cut platforms. The coastline between Hawera and Wanganui is currently retrogressive along much of its length, except for rare areas where small foredunes protect the cliff base from direct wave attack eg., immediately west of the Patea River mouth. Fleming (1953 p.22) reported a recession rate of approximately 1.5 m/year south west of the Mowhanau Stream in the period 1876-1893. However, more recently, an extensive survey of historical records by Gibb (1978) indicates an average recession rate of 0.7 m/year between Hawera and Wanganui over the past 100 years (mean of 15 localities excluding Castlecliff Beach and the Patea River mouth, both of which have been affected by harbour works and have reportedly prograded at annual rates of 5 m and 1 m respectively (Gibb 1978)). Such high rates of coastal recession would appear to be caused by exposure of soft sediments to moderate to high energy wave climates characteristic of much of the southwestern North Island (see Pickrill & Mitchell 1979).

During the winters of 1976 and 1978 I observed widespread retreat of the coastal cliffs via large collapses of relatively unconsolidated sediments. In March 1978 four people were killed near
Hawera by a single large collapse. These collapses which produce large talus cones of debris at the foot of the cliffs (Plate 2.7) are similar to those described by McLean & Davidson (1968) on the east coast of the North Island, and appear to be the dominant mode of coastal recession. Removal of fallen material by wave action was usually complete within 2-3 months of its fall, particularly when wave energies were persistently high. Wave rider buoy records on the Maui gas platform in the period March 1976 to December 1977 (Pickrill & Mitchell 1979) indicate no pronounced seasonality in offshore wave energies during this time. Similarly, a study of five western Taranaki beaches by Matthews (1979) in the period February 1976 to May 1977 showed only a weak seasonality in beach volume changes. The inevitable conclusion from such studies is that storm events causing coastal change in Taranaki are not markedly seasonal, and that day to day climatic events are likely to be the dominant control on wave climate and coastal recession.

Modern theories of cliff recession and platform genesis fall into two broad groups:

1. those in which subaerial erosion and weathering are the dominant cause of cliff recession
2. those in which wave action is dominant.

The South Taranaki coastline, with its soft sediments and moderate to high wave energies falls into the latter category. In California, where coastal morphology and processes are apparently very similar to South Taranaki, Bradley (1958) has noted that fragile, etched pyroxene grains occur in large numbers below a water depth of 30 feet (9 m) but are virtually absent in sediments from shallower depths. Thus he has argued that marine abrasion is only occurring down to a depth of 30 feet (9 m). Based on the geometry of wave cut platforms in the same area, Bradley (1958) was able to conclude that platforms wider than one third of a mile (500 m), and hence extending below 9 m water depth, must therefore have been the result of a rising relative sea
level. Furthermore, theoretical considerations (Flemming 1965) and field measurements (eg. Kirk 1977, Trenhaile 1974) suggest that cliff recession decreases with increasing platform width for a stationary sea level. Consequently it seems likely that a slowly rising relative sea level would be most favourable to produce the wide (up to 4 km wide) wave cut surfaces on the South Taranaki marine terraces. The similarity between wave cut platforms on California marine terraces and modern wave cut platforms in the same area led Bradley (1958) to conclude that little platform modification occurred during a relative sea level fall.

Each South Taranaki marine terrace (Fig. 2.5, Plate 2.8) consists of a relatively thin veneer of marine sands, often with basal conglomerate, or more rarely, shell layers, deposited immediately above a subhorizontal (dip 1°) wave cut surface. The wave cut surface is an angular unconformity which truncates older Plio-Pleistocene sediments of the underlying Wanganui Basin. Locally, the underlying rocks (and some of the larger pebbles in the base of overlying marine sands) are bored by rock boring marine organisms. The marine sediments grade upwards into peats and/or a variety of other littoral and terrestrial deposits, which may include fluvial sediments, tephras, dunesand, loess and laharic debris. The total thickness of cover beds (marine and non-marine) generally increases with age of the terrace and exceeds 40 m on some of the older terrace surfaces. Cover bed thicknesses also increase west of Hawera, where the input of volcanic material increases dramatically. A degraded, and often buried sea cliff defines the inland margin of each terrace and represents the location of the shoreline at the peak of the marine transgression which cut the wave cut surface below it (Plate 2.9 and 2.10).
Terrace nomenclature is illustrated in Fig. 2.5 Definitions of various terms are as follows:

"Marine Terrace" : includes deposits (both marine and non-marine) immediately overlying and including a wave cut surface, which together form a mappable unit; bounded at its inner edge by a fossil cliff and its outer edge by a fossil or active cliff. No particular topographic form is specified, eg. Brunswick Terrace.


"Wave cut surface" : a subhorizontal erosion surface displaying either of the following:

(a) immediately overlain by marine sediments and/or marine fossils

(b) evidence of boring by marine organisms.

Synonymns : wave cut platform (Bradley & Griggs 1976), abrasion platform (Kern 1977), cut surface (Chappell 1975).

"Cover beds" : those sediments (both marine and non-marine) overlying a wave cut surface within a terrace.

"Strandline" : a wave erosional feature which represents the intersection of a wave cut surface with the base of fossil cliff.


"Terrace surface" : topographic expression of a marine terrace; may be typically, a gently sloping dissected ground surface, bounded by cliffs at its inner and outer edges.

Synonymn : terrace (Chappell 1975).

In the South Taranaki-Wanganui region, the underlying sediments which are truncated by the wave cut surfaces are usually distinguishable from the overlying marine sediments on the basis of:

(1) their greater degree of lithification

(2) their finer grainsize
(3) their higher content of micaceous minerals and lack of mafic minerals
(4) their greater dip.

The abundance of mafic minerals in the marine (terrace) sediments (Plates 2.11 and 2.12) generally decreases eastwards, reflecting increasing distance from the volcanic source. In addition, there is also a decreasing abundance of mafic minerals with increasing age of the terrace, and this may be related to the changing frequency of volcanic activity in Taranaki and/or perhaps in part to post-depositional diagenetic effects. Both these trends therefore make recognition of the wave cut surface more difficult on the older and eastern terraces of the sequence.

Fig. 2.6 summarises the sequence of events thought to accompany the development of a marine terrace in the South Taranaki-Wanganui sequence (modified after Bradley & Griggs 1976).

Although the northern-most tip of the South Island (Cape Farewell) lies some 160 km southwest of Hawera, the shallow depth of water between (120 m) means that the North and South Islands would have been joined at the height of the last glaciation. In addition, there are two more important implications (for this study) arising from the extreme width and gentle slope of the continental shelf. Firstly, any change in the height of relative sea level will be accompanied by relatively large horizontal shoreline displacements, resulting in wide terraces and enhanced prospects of recording even small fluctuations in sea level. Secondly, tectonic tilting normal to the coastline, and hydroisostatic deformation owing to changing water loads, are likely to be important components of the total deformation of the terrace sequence. This will complicate the task of separating tectonic and eustatic effects in the terrace record.
In many places along the South Taranaki-Wanganui coast, dunefields extend inland for several kilometres at an angle to the present coastline (Plate 2.13). Towards the west these dunesands become increasingly rich in mafic minerals, and near Waverley for example are being mined for their iron content. Most of these dunefields were active during the Holocene (Cowie 1963, Fleming 1953), but are now stabilised, although some remobilisation has been caused by Maori and European settlement. Remnants of older dunefields are present further inland, and some have been buried by thick deposits of volcanic ashes. Milne (1973a) has argued that major dune advances probably only occurred during interglacial or interstadial periods; however, both Cowie (1963) and Neall (pers. comm. 1978) report that the Aokoutere Ash which is dated at about 20,000 years BP is interbedded with dunesands at some localities, and that sand movement therefore also occurred during the last glaciation.

2.6 CLIMATE

The North Island of New Zealand lies just to the south of the subtropical high pressure belt, in the zone of mid-latitude westerlies, and is therefore within the zone of interaction between subtropical and polar air masses. In summer, the mean position of the subtropical highs moves far enough south (36°) so that the North Island comes under the influence of relatively dry, subtropical air (Garnier 1958, Tomlinson 1976). In winter, on the other hand, the mean position of the subtropical highs lies well to the north of New Zealand (26°), and the whole of the country lies within the region of dominant westerlies (Garnier 1958, Tomlinson 1976). Although major depression centres, which form on the Polar Front, pass to the south, the northward extension and subsequent modification of their associated cold fronts strongly influences the day to day weather in New Zealand, particularly in winter (Garnier 1958).
Not only is the climate of New Zealand the result of its latitudinal position, but also its position in a wide expanse of ocean. This ensures that firstly temperatures are moderated, particularly on the west (windward) coasts of both islands (Tomlinson 1976), and that secondly winds bring ample moisture with them. The orographic modification of these maritime westerlies is also important. The topographic dependence of both rainfall and temperature in the Taranaki region is clearly seen in Figs. 2.7 and 2.8. Furthermore, Gamier (1958) has pointed out that the high mountains of New Zealand not only encourage cloud formation and rainfall owing to orographic uplift, but also result in mixing of the atmosphere to produce clear skies and plentiful sunshine (1800 hours/year in most places except the high mountains).

The distribution of mean annual precipitation in Taranaki is shown in Fig. 2.7. Rainfall is generally higher in North Taranaki, which is on the windward side of Mt Egmont, than it is in South Taranaki. Rainfall totals increase markedly with altitude, particularly on Mt Egmont, where at higher elevations (above 900 m), mean annual totals exceed 7000 mm (Coulter 1976). On the terraces themselves a slight increase in precipitation with increasing elevation is also observed. The driest part of the terrace sequence, just west of Wanganui, receives slightly less than 1000 mm, while in the wettest area just south of Eltham mean annual rainfall is nearly 1600 mm. To the southeast of Wanganui rainfall drops below 900 mm in some areas.

Typically, rainfall is spread evenly throughout the year, with a slight winter maximum and summer minimum, particularly in the west. In the eastern part of the region, the maximum may be in spring, and the minimum in autumn. The number of raindays with greater than 1 mm rainfall is generally between 100 and 150 per annum.
(Thomlinson 1976), although up to 200 are experienced on the higher slopes of Mt Egmont (Coulter 1976). Despite the relatively low number of raindays, average rainfall intensities tend to be low. Drizzly weather is the rule rather than the exception in Taranaki, and heavy rain is only likely to occur from deep depressions or slow moving troughs (Maunder & Browne 1972). Most places can expect to receive 50-150 mm in a day at least once every two years (Robertson 1963), although daily rainfalls of over 400 mm have been recorded at Dawson Falls (945 m) and Stratford Mountain House (846 m) on the higher slopes of Mt Egmont (Coulter 1976). Despite the relatively high mean annual rainfall over most of the region, the mean number of hours of bright sunshine per annum is quite high (1800-2200) or near average for much of New Zealand (Thomlinson 1976).

In the western half of the region, average monthly rainfall exceeds potential evapotranspiration throughout the year; however, from Patea eastwards, in the drier areas, average monthly rainfall may be less than potential evapotranspiration in summer and early autumn months (Garnier 1958, Maunder & Browne 1972). Mean monthly air humidity at 9 am averages between 70 and 90% throughout the year, with a maximum in winter and a minimum in summer (N.Z. Meteorological Service 1973b), although once again it is generally higher on the upper slopes of Mt Egmont.

Mean annual air temperatures tend to be between 10 and 15°C although at higher altitudes they fall below 10°C (Fig. 2.7, 2.9). Based on available statistics for Taranaki (N.Z. Meteorological Service 1978) the mean lapse rate is slightly less than 6°C/1000 m or 1°C/176 m (see Fig. 2.8). On the terrace sequence, mean annual air temperatures vary from 12 to 14°C and decrease slightly with altitude.
Snow occurs rarely, except at higher altitudes on Mt Egmont, where snow may remain through summer in some years. In the winter of 1976, snow fell at Ararata, just inland of Hawera, down to an altitude of about 150 m; however, local residents report that this is unlikely to occur more than 2 or 3 times a decade. The frequency of frosts increases inland from the coast and with increasing altitude; near the coast frosts are generally only experienced once or twice a year (N.Z. Meteorological Service 1973b).

Although prevailing winds are from the westerly quarter in exposed places, orographic modification is locally important. For example, Mt Egmont substantially modifies the wind pattern to the east: Stratford has a predominance of winds from north and south (Fig. 2.7). In general, to the north of Mt Egmont winds are dominantly from the west and southwest, while to the South of Mt Egmont, they are dominantly from the west and northwest (Thomlinson 1976). Average wind speed increases markedly with altitude, and probably exceeds 40 km/hour near the summit of Mt Egmont (Coulter 1976).

In summary, Maunder & Browne (1972 p.13) described the climate of Wanganui as "a reasonably pleasant climate, with few notable extremes". This description is probably true for much of Taranaki, except perhaps for the upper slopes of Mt Egmont.

2.7 VEGETATION

Little natural vegetation remains in the Taranaki lowlands, owing to extensive clearance by both Maori and European settlers, although scattered small remnants of native forest do occur in those
steeper gullies which are unsuitable for grazing or agriculture (Plate 6.8). However, substantial areas of forest remain on the upper slopes of Mt Egmont (Plate 2.1) and the Pouakai and Kaitake Ranges (Plate 2.2), particularly within the Egmont National Park. Natural forest is also still present in the more rugged inland areas to the north and east of the lowlands (Plates 2.3 and 2.4).

Even at the time of European settlement, the coastal region had largely been cleared of forest by the Maoris, and was covered with fern and scrub for several kilometres inland (Dieffenbach 1843). Previously the forest would have graded into scrub forms, comprised of hardwoods with occasional podocarps in these coastal areas (Wendelken 1976, McGlone pers. comm. 1979). On the other hand, Wendelken (1976) also pointed out that the podocarp-hardwood-beech forests of the inland area (Fig. 2.9) were probably not much more extensive in 1840 than they are today. Elsewhere, except at high altitudes (above 1200 m) and in swamps, there would have been tall, dense forest. Figure 2.9 shows the estimated distribution of major vegetation classes at the time of European settlement (c.1840).

The most widespread forest types were the lowland podocarp-hardwood forests, which contained *Dacridium cupressinum* (rimu), *Beilschmiedia tawa* (tawa), *Metrosideros robusta* (rata) and *Weinmannia racemosa* (kamahi) as dominants (Dieffenbach 1843, Godley 1976). Where soil fertility was low, *Beilschmiedia* was usually replaced by *Weinmannia* (eg. on the western and higher slopes of Mt Egmont (Druce 1976), or by *Nothofagus* or beech (eg. on the inland ridges (Cockayne 1928)). *Podocarpus spicatus* (matai) was also present, but not as common as the above mentioned species, and was probably confined to the most fertile, better drained soils (McGlone 1979 pers. comm.). In swampy areas *Podocarpus dacrydioides* (kahikatea or white
pine) was probably dominant (Godley 1976). A rich hardwood tree and shrub flora, with abundant tree ferns, is characteristically associated with the lowland prododarp-hardwood forests (Godley 1976).

With increasing altitude in the Egmont National Park, Beilschmiedia, Dacridium and Metrosideros become less abundant (Druce 1976). As a consequence of this decline, Weinmannia reaches its peak abundance between 760 and 840 m (Druce 1976, McGlone 1979 pers.comm.). Above 900 m, Libocedrus bidwillii (kaikawaka or mountain cedar) and Podocarpus hallii (Hall's totara) become the dominants in an upland podocarp-hardwood forest, which still contains Weinmannia below 1000 m (Druce 1976). Above 1100 m the forest gradually gives way to a subalpine scrub community, and above about 1400 m, is in turn replaced by alpine grassland (Druce 1976, McGlone pers.comm.1979). The vegetation zonation on Mt Egmont is clearly seen in Plate 2.1.

Inland from the Taranaki lowlands Nothofagus forest is found on the ridges and spurs of the heavily dissected upland. In these podocarp-hardwood-beech forests, Beilschmiedia, Weinmannia, Dacridium and Podocarpus ferrigineus (micro) are common, together with one of the several beech species (McGlone pers.comm.1979). Nothofagus fusca (red beech) is found throughout the area. In the north, N. truncata (hard beech) is common, while in the south N. solandri var. solandri (black beech) is common (Wendelken 1976). N. menziesii (silver beech) is rare. The absence of Nothofagus from the Egmont National Park area is conspicuous (Druce 1976).
3.1 INTRODUCTION

As early as 1815, the French physicist Jean Baptiste Biot observed that certain naturally occurring organic compounds were able to rotate plane-polarised light either clockwise or anticlockwise when in solution (Fieser & Fieser 1961, p.67). However, it was not until 1874 that Van't Hoff and Le Bel independently deduced that in biological molecules this optical activity was related to the tetrahedral spatial arrangement of bonds around the carbon atom (Fieser & Fieser 1961, p.73). When four different groups are present the carbon atom to which they are attached is termed "asymmetric" or "a centre of asymmetry". This enables a molecule containing an asymmetric centre to exist in two stereomeric forms which are mirror images of each other. These two forms are termed enantiomers and each will rotate plane-polarised light in opposite directions. Apart from this property they are physically and chemically indistinguishable. When more than one centre of asymmetry is present, the molecules are termed "diastereomeric". In contrast to enantiomers, diastereomers are not mirror images of each other, and usually have substantially different physical properties (Roberts & Caserio 1964).

Amino acids, or \(\alpha\)-amino carboxylic acids, are optically active molecules characterised by a centre of asymmetry at the \(\alpha\)-carbon position (the carbon atom to which the carboxyl group (-COOH) is attached is referred to as the \(\alpha\)-carbon atom; other carbons are numbered \(\beta, \gamma, \ldots\) successively from the \(\alpha\)-carbon). An hydrogen atom (-H), an amino group (-NH\(_2\)) and an -R group are also attached to the \(\alpha\)-carbon (see Fig. 3.1A), where the -R group is different for each of the 20 or so naturally occurring protein amino acids which
make up the basic structure of all living organisms (see Fig. 3.2). The two stereoisomers of any amino acid with a single centre of asymmetry are, by convention, named the L- and D- enantiomers (see Fig. 3.1A) with respect to the assumed configuration of D(+) glyceraldehyde; however, it should be noted that while these names are derived from the terms levorotatory (rotating the plane of polarised light to the left or counterclockwise) and dextrorotatory (rotation to the right), owing to the naming convention, not all L-amino acids are in actual fact levorotatory (current practice is to use (+) and (-) signs to distinguish dextro- and levorotatory respectively).

In living organisms, nearly all the amino acids are L-isomer*, although the D-isomer is occasionally present (Meister 1965, Corrigan 1969). However, in normal laboratory syntheses, a racemic mixture of equal amounts of each isomer is produced i.e. optically inactive, and plane polarised light passes through the solution without rotation. For those amino acids with two centres of asymmetry (isoleucine, threonine and hydroxylysine), four isomers can theoretically exist, although in nature it appears as though only two are normally present (see Fig. 3.1B for example). Dungworth (1976) has suggested that the very high activation energy needed for inversion at the α-carbon atom may prevent the formation of D-isoleucine and L-alloisoleucine in natural environments.

According to Schroeder & Bada (1976), when an organism dies, its proteins and amino acids may undergo various chemical and biological changes which include:

(a) hydrolysis of peptide bonds, releasing smaller peptide units, and eventually free amino acids (see Fig. 3.3)

(b) decomposition reactions, which vary according to which amino acid is involved

* The simplest protein amino acid, glycine, and the two non-protein amino acids β-alanine and γ-aminobutyric acid (Fig. 3.2) have no centre of asymmetry and hence only exist in one form.
(c) degradation by microorganisms, which use the amino acids as a food source

(d) condensation reactions with other molecules, to produce complex polymers such as humic acids

(e) racemisation, which eventually produces an equilibrium mixture of D- and L-isomers for each amino acid. (For amino acids with more than one centre of asymmetry this is more correctly termed epimerisation).

The racemisation reaction forms the basis for a relatively new chemical dating technique, and is used here to estimate the age of fossil wood specimens which are beyond the range of conventional C\textsuperscript{14} dating methods (ie. older than c.40Ka BP). The rate at which racemisation proceeds varies from amino acid to amino acid (eg. aspartic acid has the fastest rate and valine has the slowest rate), consequently individual amino acids must be separated from the fossil matrix prior to analysis. This separation is normally accomplished by:

(a) Hydrolysis of the sample in a strong acid (eg. 6NHC\textsubscript{1}) to release individual amino acids from their protein matrix.

(b) Separation of the individual amino acids on a commercially available ion-exchange amino acid analyser, to measure their concentrations. A typical auto-analyser output (chromatogram) is shown in Fig. 3.3B, and this data will be referred to here as an amino acid spectrum. Note here that the relative concentration of D and L-isomers (extent of racemisation) can only be determined in this way for the diastereomeric amino acids eg. the two naturally occurring stereoisomers of isoleucine (L-isoleucine and D-alloisoleucine - see Fig. 3.1B) appear as separate peaks on the chromatogram. Amino acids that have only a single centre of asymmetry appear as a single peak, which represents a mixture of D- and L-isomers; their separation requires much more complicated techniques and apparatus. Partly for this reason, only the extent of racemisation (epimerisation) of isoleucine is used in this thesis for age determinations.

The amino acid racemisation dating method and its application are described fully in the following sections.

3.2 BACKGROUND

Although the racemisation of amino acids under extreme conditions of pH and temperature has been known for some time, it has only recently been documented that firstly the reaction will occur
under the more "normal" conditions found in nature, and secondly that amino acids can survive for long periods of time in natural environments, given favourable conditions.

Hare & Abelson (1967) were the first to identify racemisation of fossil proteins, when they reported that racemisation had proceeded to equilibrium for several amino acids in fossil shells of Miocene age. The following year, Hare & Mitterer (1968) first suggested that the racemisation reaction could be used for dating fossil materials by using the increase in $^{D/L}$ ratio with increasing time (i.e. increasing racemisation) as an indication of the age of the material. Bada (1972a) studied the kinetics of racemisation of various free amino acids in solutions over the pH range 1-13, and showed that racemisation proceeded at values of neutral pH at rates little different than in acidic or basic solutions. Since this time, much work has been carried out on the extent of racemisation in a wide variety of organic materials, both living and fossil, including:

(a) bone (Bada 1972b; Bada & Protsch 1973; Bada & Helfman 1975; Bada et al. 1974; Bischoff & Childers 1979)

(b) molluscs (Mitterer 1974, 1975; Andrews & Miller 1976; Wehmiller & Belknap 1978; Kvenvolden et al. 1979)

(c) marine sediments (Bada et al. 1970; Wehmiller & Hare 1971; Bada & Schroeder 1972; Bada & Man 1980)

(d) lacustrine sediments (Schroeder & Bada 1978)

(e) foraminifera (Kvenvolden et al. 1973; King & Neville 1977)

(f) meteorites (Kvenvolden et al. 1970; Lawless 1973)

(g) wood (Lee et al. 1976; Engel et al. 1977)

(h) corals (Wehmiller et al. 1976)

(i) soils (Pollock et al. 1977)

3.3 KINETICS AND MECHANISM OF THE AMINO ACID RACEMISATION REACTION

The general amino acid racemisation process can be described by the first order reaction:

$$\frac{K_L}{K_D}$$

$\text{L-amino acid} \xrightleftharpoons{K_D} \text{D-amino acid}$
where $K_L$ and $K_D$ are the rate constants for the forward and reverse reactions. At any one time, the rate of disappearance of the L-amino acid depends not only on the amount of L-amino acid being converted to D-amino acid, but also on the amount of D-amino acid being converted to L-amino acid, i.e., the difference between the processes of formation and disappearance of the L-amino acid. This can be written:

$$-\frac{d[L]}{dt} = K_L[L] - K_D[D] \quad (1)$$

where $[L]$ and $[D]$ are concentrations of the L- and D-isomers, respectively. Integration over the age of the sample $(t)$ gives:

$$\log \left( \frac{1+(D/L)}{1-K'(D/L)} \right) - \log \left( \frac{1+(D/L)}{1-K'(D/L)} \right)_{t=0} = (1+K').K_L.t \quad (2)$$

where $K' = K_D/K_L = 1/Keq$ is the reciprocal of the equilibrium constant (Bada & Shroeder 1972).

The second term on the left hand side of equation (2) will equal zero if there is no D-isomer present at time $t = 0$. In practice, however, this is rarely the case owing either to small amounts of D-isomer present in the sample at the time the organism died, or to the small amount of racemisation which may occur during sample preparation. (Acid hydrolysis at 105°C in 6N HCl is the standard laboratory procedure for breaking up the protein bound amino acids into free forms for subsequent analysis. A small amount of racemisation may be induced during this procedure).
For amino acids with only one centre of asymmetry, $K_L/K_D$ is exactly equal to one, but for those with more than one asymmetric centre, $K_L/K_D$ differs slightly from one, reflecting the greater stability of one of the isomers relative to the other. For example, in the case of the epimerisation reaction $L$-isoleucine $\neq D$-alloisoleucine, the equilibrium constant has variously been determined to be between 1.25 and 1.40 in a wide variety of materials (Williams & Smith 1977) reflecting the greater stability of alloisoleucine relative to isoleucine.

Assuming that the racemisation reaction is described by the simple first order kinetic model above, by measuring the $D/L$ ratio in a fossil, and estimating both the rate constant ($K_L$) and the equilibrium constant ($K_{eq}$), equation (2) may be used to calculate the age ($t$) of the sample. The rate constant $K_L$ may be calculated using either:

1. **The calibration method**, by analysis of samples of known age (Bada & Protsch 1973).  

or

2. **The kinetic method**, by extrapolation from elevated temperature studies (eg. Mitterer 1974) using the behaviour of $K(T)$ as described by the Arrhenius equation (Eqn (3) p.49), where $T$ is absolute temperature.

The suggested mechanism (Shroeder & Bada 1976) for the amino acid racemisation reaction involves removal of a proton attached to the $\alpha$-carbon atom, leaving behind a planar carbanion (ie the three remaining bonds all lie in the same plane, whereas previously the four bonds were tetrahedrally arranged around the $\alpha$-carbon see Fig. 3.1A.) Since the readdition of the proton can occur from either side of the molecule with equal probability, there is an equal probability of

* Variation in the equilibrium constant ($K_{eq}$) is partly related to whether the amino acids are bound or in the free state, and if bound, the type of matrix in which they are bound. $K_{eq}$ (iso/alloisoleucine) of 0.86, determined for wood by Lee et al. (1976) is somewhat anomalous, and appears in some way to reflect the effects of the organic matrix on the epimerisation reaction.
forming either the L- or D- enantiomer, and eventually this leads to an equilibrium mixture of the two. Removal of the proton may theoretically be initiated by any nucleophile species; in aqueous solution the nucleophile is apparently the OH-ion (Schroeder & Bada 1976). In fact the absence of reaction in heated samples of anhydrous protein (Hare 1974) and the anomalously low racemisation rates in fossils from the La Brea tar pits in California (Kvenvolden & Peterson 1973), indicate that the presence of water may be fundamental to the racemisation reaction.

In distinguishing the three major groups of amino acids found in fossil materials:

(a) protein bound progressive natural hydrolysis
(b) peptide bound and breakdown of protein
(c) free structure

Bada & Shroeder (1972) also noted that the rate of racemisation increased with progressive hydrolysis (ie. from (a) to (c); see also Fig. 3.3). Similar observations had previously led Wehmiller & Hare (1971) to conclude that a significant proportion of racemisation must occur at the time of hydrolysis; however, as pointed out by Bada & Schroeder (1972), the hydrolysis and racemisation reactions involve entirely different carbon centres, and therefore the two reactions are likely to be completely independent. Instead, Bada & Schroeder (1972) postulated that the increased rate of racemisation accompanying progressive hydrolysis, was more likely to be the result of metal ion catalysis after the amino acids had been released by natural hydrolysis. This latter mechanism appears much more tenable in view of the fact that trace metal ion catalysis of free amino acids in aqueous solution has been well documented by Bada & Schroeder (1975), while for protein and peptide bound amino acids, the reduction in the
number of chelatable groups, owing to increased peptide bonding, theoretically makes metal ion catalysis much less significant (Davies & Treloar 1977).

Despite some uncertainty regarding the precise mechanism responsible, it is clear that a linear kinetic model is inappropriate (Wehmiller & Hare 1971; Bada & Schroeder 1972; Kvenvolden et al. 1973) to describe the racemisation reaction in those cases where:

(a) hydrolytic and catalytic reactions have occurred to any significant extent, or

(b) those amino acids which have been affected by hydrolysis/catalysis have not been removed from the sample prior to analysis.

Thus early attempts to date deep sea sediments by Bada et al. (1970) were unsuccessful, because a significant proportion of the amino acids in the sediments had undergone hydrolysis/catalysis, and no attempt was made to exclude them prior to analysis.

In most recent studies, removal of free and peptide bound amino acids, which may have undergone metal ion catalysis, has been carried out prior to acid hydrolysis by treatment with dilute HCl at room temperature ie. only the protein bound fraction of the total amino acids in the sample is analysed. Engel et al. (1977) claimed that a 1.5 NHCl preliminary wash removed more than 99% of the nonbound amino acid fraction in fossil wood samples which they analysed.

3.4 ENVIRONMENTAL FACTORS INFLUENCING THE RATE OF RACEMISATION

Unfortunately for earth scientists interested in using the amino acid racemisation reaction as a geochronologic tool, the amount and rate of racemisation are dependent not only on elapsed time, but also on the physical and chemical environment in which the amino acids are preserved. For example, the well known dependence of the rate constant \( K_L \) on temperature may be expressed using the Arrhenius equation as:

\[ K_L = A \exp \left( -\frac{E}{RT} \right) \] (3)
where $T = \text{absolute temperature (°K)}$

$A = \text{constant (hr}^{-1})$

$E = \text{activation energy (cal.mol}^{-1})$

$R = \text{gas constant (cal. °C}^{-1}.\text{mol}^{-1})$

The effect of temperature on the racemisation reaction is perhaps best illustrated in terms of reaction half-lives. For a simple first order irreversible chemical reaction, the half-life is defined as the time taken for half of the starting material to disappear, and is independent of the initial concentration (cf. radioactive decay). However, in the case of the reversible racemisation reaction, the half-life is defined as the time taken for the reaction to move from its initial "non-equilibrium position" towards equilibrium by a fraction of half of the original distance from equilibrium. This may be written as:

$$\frac{Xe - X'}{Xe} = \frac{1}{2} = \text{fraction of completion of reaction} \quad (4)$$

where $Xe = \text{equilibrium ratio } \frac{D}{D+L}$

$X' = \frac{D}{D+L} \text{ at any time } t.$

Rewriting equation (2) in a different, but equivalent form (Wehmler & Hare 1971), we have:

$$\frac{Xe - X'}{Xe} = \exp \left[ -\left(1+\frac{K_L}{K_D}\right) K_L \cdot t \right] \quad (5)$$
and combining equations (4) and (5), we obtain an expression for reaction half-life \( \tau \):

\[
\tau = \log_e 2/(1+K')KL
\]

From equations (2) and (6), an expression for \( D/L \) ratio at time may be derived:

\[
(D/L)_\tau = Keq/(Keq+2)
\]

Thus for enantiomeric amino acids with \( Keq = 1 \), \( (D/L)_\tau = 0.33 \); while for isoleucine/alloisoleucine epimerisation in wood with \( Keq = 0.86 \), \( (D/L)_\tau = 0.30 \).

Table 3.1 illustrates the effect of temperature on racemisation rates using the half-life expression for isoleucine/alloisoleucine epimerisation in wood, calculated for various temperatures. Kinetic data are from Lee et al. (1976). It is clear from Table 3.1 that small differences in temperature, produce correspondingly large differences in half-life (e.g. 1°C difference in temperature produces a 20% difference in half-life). Thus applying this to a field situation, two samples of equivalent age, but with slightly different temperature histories, can potentially yield quite different apparent ages as a result of their differing temperature histories. This means that a detailed knowledge of the temperature histories of any samples we wish to date, is critical to the successful application of the method. As Bada & Schroeder (1976) have concluded, the general agreement between racemisation paleotemperatures* and those deduced by other methods may be taken to indicate that temperature is the major environmental factor influencing racemisation rates.

In addition to temperature, several other factors exert a strong influence on the rate of racemisation over time. These have been listed in approximate order of importance in Table 3.2.

*An estimate of "effective diagenetic temperature", which will be higher than the "average diagenetic temperature" owing to the exponential term in the Arrhenius equation, may be obtained from samples of known age using equations (2) and (3).
Perhaps the next most important factor after temperature is the composition of the material in which the amino acids are preserved (matrix material). Figure 3.4 clearly indicates that the rate of racemisation of protein bound isoleucine in shell or bone matrices is much faster than for those bound in wood. The reasons for this variation remain unclear, although Bada & Schroeder (1972) have suggested that catalysis by trace metal ions within the carbonate matrix might explain the faster rate observed in calcareous deep sea sediments. However, as outlined previously, such catalysis has been demonstrated for free amino acids in aqueous solution but has yet to be demonstrated, and is theoretically unlikely, for protein bound amino acids. What is clear, is that the variation in racemisation rate with matrix material is sufficient to prevent extrapolation of kinetic data from one matrix material to another. In other words, $K_L$ should be determined on matrix material as similar as possible to that we wish to date. Furthermore, such a constraint should apparently be even more rigidly enforced so as to include only matrix material of a single species. For example, Mitterer (1974), Miller & Andrews (1977) and Kvenvolden et al. (1979) have all demonstrated a marked variation in extent of racemisation from different species of fossil mollusc which were of apparently identical ages (although no mention is made of the possible effects or presence of diagenesis). King & Neville (1977) have also demonstrated that the rate of racemisation in fossil planktonic foraminifera from deep sea sediments is strongly species dependent. In addition, Engel et al. (1978), in a study of uniformly preserved plants from an 11,000 year old packrat (Neotoma spp) midden, have shown that the extent of aspartic acid racemisation varied not only among different taxa, but also apparently with anatomical position within a single species (e.g. twigs versus pods). The inevitable conclusion from all these studies, plus consideration of the effects of hydrolysis and trace metal ion catalysis (discussed
previously), is that the microstructural environment of individual amino acids in their matrix materials is a very significant factor in determining their rates of racemisation (Engel et al., 1978, p.1560 concluded from their study that: "It is probable that some types of complex physiochemical relationships existing between bound aspartic acid and its host lignin-cellulose, etc, matrix and/or the relative position of aspartic acid in the amino acid sequences of peptides/proteins in different taxa and anatomical sites control bound aspartic acid racemisation rates").

Of the other factors in Table 3.2, perhaps the most interesting has been the recognition that certain clay minerals are able to catalyse the racemisation reaction of free amino acids in aqueous solution. Frenkel & Heller-Kallai (1977) reported the catalytic effect of montmorillonite on racemisation rates and supported the conclusions of Jackson (1971) and Thomson & Tsunashima (1973) in suggesting that the greater the surface acidity of the clay mineral, the greater the degree of induced racemisation. Furthermore, the acidity of the clay mineral depends not only on the mineral itself, but also on the exchangeable cations present, so that the presence of certain clay minerals might be a suitable source of trace metal ion catalysis.

### 3.5 Contamination

As Williams & Smith (1977, p.133) have stated: "Possibly one of the greatest sources of error in applying amino acid racemization data to dating is in the determination of the "true" D/L ratio of any amino acid because of the possibility of contamination by more recent L-amino acids". Figure 3.5 shows several potential sources of this contamination, each of which is capable of altering D/L ratios to such an extent that dating becomes virtually impossible.
3.5.1 SOURCES OF CONTAMINATION

Considering firstly the possibility of diagenetic formation of amino acids, it is surprising to note that this aspect of contamination has received very little attention in the literature. Pyrolysis experiments by Vallentyne (1964, 1969) have shown that certain amino acids become unstable when heated strongly, and that these apparently decompose to a range of products which includes several of the other more stable amino acids. Furthermore, the pattern of stability in these laboratory experiments agrees quite well with spectra of those amino acids found in very old fossils (Abelson 1954, Hare & Mitterer 1968; King and Hare 1972). Recent work by Bada et al. (1978) has demonstrated that serine and threonine, both of which are thermally unstable and present in only minor amounts in very old fossils, show a progressive decrease with increasing age in foraminifera from dated deep sea cores, and that their decomposition products include alamine, glycine and α-amino-n-butyric acid. Similarly, Petit (1974) has noted that deamination of asparagine and glutamine may occur during acid hydrolysis, to produce aspartic and glutamic acids respectively. Thus it is clear that since each amino acid has its own characteristic rate of racemisation, decomposition of one amino acid to another (as documented above) during the course of the racemisation process, will result in altered D/L ratios and produce spurious ages. Unfortunately it is also clear that since amino acids produced by diagenesis are in most cases indistinguishable from the amino acids originally in a fossil, little is known of their contribution to the overall amino acid concentrations we measure (Williams & Smith 1977, p.132).

Another source of contaminating amino acids is via their removal or addition by groundwater percolating through the sample. For example, this form of contamination has been postulated to account
for the presence of predominantly L-amino acids in Precambrian Fig Tree Chert (Kvenvolden et al. 1969; Oro et al. 1971). Similarly, the lack of concordance between racemisation ages and U/Th dates on fossil corals has been explained by Wehmiller et al. (1976) as either the result of extensive leaching of free amino acids from the fossils, or groundwater contamination by "young" amino acids. However, as Bada et al. (1973) point out, it should still be possible to calculate a minimum age, in those cases where contamination has not been too great, by assuming that either all additions of contaminating amino acids have been in the L-form, or that all losses have been free amino acids with higher D/L ratios than those remaining in the bound form.

Additions from bacteria and other microorganisms are a third possible source of contaminating amino acids in fossil materials although once again very little is known of their relative importance. Certainly some microorganisms are known to concentrate certain D-amino acids in their cell walls (Corrigan 1969), while Pollock et al. (1977) have detected significant amounts of D-alanine, aspartic and glutamic acids in contemporary soils which they attribute to this source. Thus it is likely that incorporation of microorganisms within a fossil sample will produce a greater apparent age.

The fourth major source of contamination is human contamination, which may occur during sample collection and/or preparation. When dealing with fossil samples, in which the concentration of amino acids is typically of the order of micromoles to nanomoles per gram of sample, even the contribution of amino acids from a single fingerprint (Hamilton 1965) will be sufficient to substantially alter the original amino acid spectrum. Similarly, trace amounts of amino acids in commercially available reagents may contaminate the sample during laboratory preparation.
3.5.2 PROCEDURES FOR DETECTION AND MINIMISATION OF CONTAMINATION

Schroeder (1975) showed that the non-protein amino acids β-alanine and α-aminobutyric acid, were abundant in marine sediments, but completely absent in thoroughly cleaned foraminifera isolated from the same sediments. Thus he argued that these two amino acids, which are apparently formed by the enzymatic decarboxylation of aspartic and glutamic acids by microorganisms in the sediments, could be used as tracers to indicate the effectiveness of laboratory cleaning procedures in isolating the foraminifera. He listed the following as standard laboratory cleaning procedures:

(a) direct visual inspection to determine the presence or absence of clay minerals,
(b) repeated ultrasonication/decantation cycles until no further change in amino acid spectrum and/or D/L ratios is observed*,
(c) observation that amino acid enantiomeric equilibrium has been reached in very old samples,

and suggested that the efficiency of these could be verified by checking for the presence of the two non-protein amino acids. Presumably similar tests could also be applied to samples from terrestrial environments.

Several major sources of contamination can apparently be eliminated by extracting free amino acids with dilute HCl at room temperature prior to the hydrolysis step. This procedure has been adopted by several workers (eg. Kvenvolden et al. 1979; Mitterer 1975; Masters & Bada 1977; Engel et al. 1978), and has the important consequence of isolating the bound amino acids so that they are free of their hydrolysis products. Since the activities of microorganisms probably do not affect the D/L ratios or the amino acid spectrum of

* Katz & Man (1979) have recommended a maximum ultrasonication time of 30 minutes, owing to changes induced in the amino acid content of samples subjected to ultrasonication over long periods, presumably owing to the energy input which induces diagenetic changes).
bound amino acids, but rather, break down the protein structure to produce free amino acids, any contamination from this source could be removed by HCl pretreatment. Similarly, contaminating free amino acids from groundwater or human sources could also be removed. On the other hand, the incorporation of whole microorganisms within the sample appears likely to contribute bound amino acids which will not be removed by such pretreatment. Thus, although Wyckoff (1972) has suggested that bacteria have sufficiently different amino acid spectra for this source of contamination to be recognised, it seems likely that say a 5% contribution from this source might go undetected, and yet still be sufficient to significantly alter the measured $^{D/L}$ ratios.

Perhaps the most promising test for contamination has been suggested by Bada et al. (1973) and Dungworth et al. (1976). They have shown that the well documented variation in rate constants for individual amino acids (which may vary by up to an order of magnitude for different amino acids under the same environmental conditions), follows the same order:

$$\text{ASP} > \text{ALA} = \text{GLU} > \text{ILE} > \text{LEU} > \text{VAL}$$

in a wide variety of materials and that deviations from this order might be taken to indicate contamination. These differences in racemisation rate for various amino acids have been explained in terms of the increasing electron-withdrawing, and resonance stabilising capacities of the R-substituents attached to the $\alpha$-carbon, which can cause a corresponding increase in the rate of $^{D/L}$ interconversion via easier removal of the proton (Bada 1972b; Bada et al. 1973).
3.6. ANALYSIS OF AMINO ACIDS IN NEW ZEALAND FOSSIL WOODS

In this study, the usefulness of fossil woods in dating Quaternary deposits on the South Taranaki-Wanganui marine terraces was investigated using the racemisation (or more correctly, epimerisation) reaction isoleucine ≠ alloisoleucine. Approximately forty samples of fossil wood were collected from a variety of stratigraphically important sites, whose ages apparently ranged from less than 40Ka to greater than 200Ka. Twelve of the apparent youngest samples were submitted for C14 dating at the ANU Radiocarbon Laboratory, while two of the samples had already been dated in New Zealand at the Institute of Nuclear Sciences. Table 3.3 lists details of all C14 dated samples whose amino acid compositions have also been determined. Full stratigraphic relationships and implications are described in subsequent chapters.

3.6.1 ANALYTICAL METHODS AND SAMPLE PREPARATION

All samples were sealed in plastic bags immediately after collection in order to minimize human contamination; however, a small amount of human handling was unfortunately unavoidable during field collection. Consequently, laboratory preparation routinely consisted of cleaning with a sterilised surgical scalpel to remove surface layers, and then repeated sonication in four times distilled H2O to remove adhering particles of extraneous matter (ultrasonication times were limited to a maximum of 15 minutes). After preliminary cleaning and visual inspection, approximately 5g of each sample was broken into small pieces and allowed to stand in 6NHCl at room temperature for 24 hours to remove free amino acids*. After 24 hours, the HCl was decanted, and the samples washed in four times distilled H2O several times to remove all trace of HCl. During preliminary laboratory cleaning, samples were only handled using sterilised plastic gloves.

* Some bound amino acids may also be removed in this step (Lee et al., 1976).
All glassware was carefully cleaned to prevent contamination from fingerprints. 10N HCl, which was sub-boiling distilled in an all glass apparatus, was diluted to 6N with 4 times distilled H₂O (distilled HCl and H₂O were provided by the Research School of Earth Sciences, ANU).

Following the above treatment to remove free amino acids, an excess of 6N HCl was added and the samples refluxed at 80°C for 24 hours. This hydrolysis was carried out at 80°C, rather than the normal temperature of 105-110°C, in attempt to minimise the amount of induced racemisation. Subsequent analyses showed the presence of few, if any peptides in the resultant hydrolysate, indicating that 80°C was sufficient to break down the bound amino acids to their free form. No attempt was made to exclude air from the hydrolysis step. Hydrolysates were then filtered, and evaporated to dryness on a rotary evaporator. No desalting was carried out. Samples were then taken up in sodium citrate buffers and analysed on a modified dual column Technion Auto-Analyser Model AAA-1 (see James 1972 for details). Amino acid concentrations were calculated manually using peak areas. Resultant amino acid spectra are listed for all samples in Appendix 1. The diastereomers L-isoleucine and D-alloisoleucine peak were resolved on the amino acid analyser, although the alloisoleucine peak was often difficult to distinguish owing to its smallness and interference from other peaks. D/L ratios were calculated where possible and are also listed in Table 3.4 as well as Appendix 1.

Free amino acid spectra were measured on 13 samples as a check of the proportion of free vs bound amino acids present.
3.6.2 INTERPRETATION OF RESULTS

The amino acid analyses listed in Appendix 1 indicate several interesting features of the amino acid composition of South Taranaki fossil woods. Firstly, the bulk of the amino acids in each sample appear to be in the bound form. In those eight samples where the total concentration of free and bound amino acids was measured, the average free/bound ratio was 6%. In addition, an attempt was made to measure the concentration of free amino acids in 5 other samples not listed in Appendix 1, all of which yielded peaks on the amino acid analyser output charts which were too small to measure. These observations were confirmed qualitatively during sample preparation and subsequent analysis by using hydrolysate colour as a guide to amino acid concentration: it was found that those hydrolysates containing very low concentrations of amino acids had pale lemon-yellow colours, and with increasing amino acid concentration the hydrolysates become darker, to yellow-brown or brown; free amino acid extracts were predominantly pale lemon-yellow, while bound extracts were invariably darker. This suggested not only that the bulk of the amino acids were in the bound form, but also suggested that the pigmentsing agents (lignins or tannins?) were closely associated with the amino acids in the wood structure and were also liberated during acid hydrolysis. On the other hand, it should be noted that samples 030A (peat) and 030B (wood) of identical age and site which were treated with 6N HCl at room temperature initially for twenty four and then for a further 61 hours (see Table 3.4) both yielded higher concentrations of "free" amino acids after the second treatment. While these concentrations were still more than an order of magnitude smaller than the concentrations of bound amino acids (BJP-030E and 030F - Table 3.4), it indicated that not all "free" amino acids were being removed by the standard 24 hour treatment. (Almost certainly the extended HCl treatment was degrading the protein structure to yield more "free" amino acids).
In general a decrease in both total and bound amino acid concentrations with increasing age of the sample might have been expected (reflecting greater decomposition). For example Lee et al. (1976) suggested that this trend, in the limited number of samples (5) which they examined, was the result of progressive natural hydrolysis of bound amino acids to free amino acids, and their subsequent removal from the wood. Similar trends have also been well documented in deep sea sediments (Bada & Man 1980) and bone (Bada et al. 1973). However, as shown in Fig. 3.7, when concentration vs D/L ratio is plotted for the South Taranaki wood samples, there is considerable scatter, suggesting that degradation has been partly controlled by other factors. This variability in decomposition is also confirmed by X-ray diffraction studies on some of the samples: samples of similar age show considerable variation in degradation of wood structure (MJ Head pers. comm. 1980). It therefore seems likely that local site variations caused by biological activity or physical factors such as redox potential may have been important in determining the preservation state of the wood samples, particularly in the critical period soon after the death of the plants.

Lee et al. (1976) have reported that with time, a decrease in the relative amounts of aspartic acid, threonine and serine occurred in their fossil woods relative to valine, while the relative amounts of isoleucine, leucine and glutamic acid remained constant. Similar results have been obtained in studies of deep sea sediments (Bada & Man 1980). Lee et al. (1976) also reported that proline became dominant in older woods. While the above observations are in accord with stability patterns observed in laboratory studies and in very old samples (see 3.5.1 Sources of Contamination), Figs. 3.8, 3.9 and 3.10 once again demonstrate considerable scatter in trends with time (D/L ratio) for the Taranaki samples. Local effects on preservation state
are again suggested as a contributing factor, although species variation almost certainly plays a part as well: unfortunately it was not possible to restrict the dated samples to a single species, nor was it possible to have all samples identified (see Table 3.5). Consequently the effects of species variation on amino acid composition and racemisation rate remain uncertain at this stage. These points raise one of the potentially limiting factors in using fossil woods for racemisation dating: the problem of identifying samples. The samples listed in Table 3.5 were kindly identified by Mr R.N. Patel, Botany Division, D.S.I.R., Christchurch; however, unlike fossil shells, for example, which can be identified readily in the field, identification of fossil woods requires considerable laboratory time and effort as well as being a skill which is little known other than to a few specialists.

Table 3.4 summarises bound alloisoleucine/isoleucine ratios and age determinations for the South Taranaki wood samples. The calibration method (Bada & Protsch 1973) of age determination was used. Samples BJP-026, 027, 028 and 029 have been assigned an age c.400Ka BP based on a c.370Ka fission track age determination from an overlying volcanic ash (see 4.4 Tephrostratigraphy). Kinetic data from Lee et al. (1976) (Keq 0.86), together with a mean D/L ratio of 0.112, were used in equation (2) to give an in situ rate constant of $2.59 \times 10^{-7} \text{yr}^{-1}$ for the terrace samples (see Table 3.4). By assuming that:

1. samples both younger and older than the calibration samples have experienced similar effective diagenetic histories,

2. other environmental factors have been relatively uniform within the terrace sequence eg. hydration, Eh, pH etc.,
3. No contamination is present, ages have been calculated for all other samples.

An initial D/L ratio of 0.01 has been assumed for all age determinations and is based on D/L ratios in C14 dated samples (Table 3.3). The dating curve, based on the above data is given in Fig. 3.11.

The above assumptions appear reasonable in the light of the following observations:

1. All samples lie within a climatically relatively uniform region (see 2.6 - Climate). In particular present mean annual air temperatures are very similar over the entire sampling region (<2°C variation - see Fig. 2.7).

2. All samples were buried at least 5m below the ground surface for the major part of their history. They would therefore have experienced only relatively long period, low amplitude temperature fluctuations, with a correspondingly damped effect on racemisation rate variations. Furthermore, although such long period, low amplitude fluctuations have been occurring throughout the Pleistocene, (a crude approximation of these is shown in Fig. 3.12), long term average temperatures experienced by samples of differing ages can be shown to remain relatively constant (see Fig. 3.12A).

3. Within the limits of experimental uncertainty, all dates are in accord with stratigraphic data. Very young samples, independently dated by C14 methods have very low or negligible D/L ratios, while the four oldest samples yield the highest, and very similar ratios. Samples of similar
stratigraphic age have, in general, similar racemisation ages. The species effect is therefore not marked.

4. The geochemical environments from which the samples are derived are probably quite similar in terms of matrix material, clay mineralogy, hydration and perhaps pH, although it has already been noted that the state of preservation is quite variable in some samples of similar age.

A more critical evaluation of the amino acid racemisation dates presented in this work must ultimately await comparison with dates derived from other methods. In particular, for the South Taranaki terraces, fission track (on tephras) and U-series (on shells) dating methods appear likely candidates. On the other hand, it is possible to say already that the dates are stratigraphically consistent and also compare favourably with age estimates which are made later in the thesis, on the basis of terrace deformation and correlation with overseas sea level curves.

It might appear at first glance, that the use of an equilibrium constant \( \text{Keq} = 0.86 \), determined by Lee et al. (1976) on North American Bristlecone Pine, could have led to spurious ages; however, the effect of \( \text{Keq} \) on calculated ages using the calibration method is quite small, even for large variations in \( \text{Keq} \) (see Fig. 3.6). In fact of much greater significance to the age determinations (and calibration curve) presented here is the uncertainty in the initial \(^{\text{D}}/^{\text{L}}\) ratio in the samples - assumed to be 0.01 on the basis of \(^{\text{C}}\text{14}\) dated samples (see Table 3.3, Fig. 3.11). Variation in age determinations, expressed as % difference from the present model, are shown in Fig. 3.13 for various initial \(^{\text{D}}/^{\text{L}}\) ratios. Quantitatively, these
deviations are much more significant, particularly in young samples, than deviation owing to variation in selection of equilibrium constant. The effect of in situ temperature variation between samples (see Table 3.1) is also shown for comparison in Fig. 3.13.

One final and interesting set of calculations which can be made using the racemisation data are those relating to paleotemperatures. As indicated previously (p.51), it is possible using samples of known age and using equations (2) and (3), to calculate "effective diagenetic temperatures" that samples have experienced since death. For example using the 400Ka calibration samples and substituting:

\[ K_L = 2.59 \times 10^{-7}\text{yr}^{-1} = 2.95 \times 10^{-11}\text{hr}^{-1} \text{ (this study)} \]
\[ A = 3.853 \times 10^{12}\text{hr}^{-1} \text{ (Lee et al 1976)} \]
\[ E = 29400 \text{ calmol}^{-1} \text{ (Lee et al 1976)} \]
\[ R = 1.987 \text{ calmol}^{-1}\text{deg}^{-1} \]

in the Arrhenius equation (equation (3)), yields an effective diagenetic temperature of 278°K (5°C) for the calibration samples. This derived temperature (a rate averaged temperature: cf. exponential term in Arrhenius equation) will be greater than the simple mean (or time averaged) temperature the samples have experienced. Therefore, such a temperature, which is some 8°C lower than present mean sea level temperature in Taranaki (see Fig. 2.8), implies a glacial-interglacial temperature difference of at least 16°C (assuming symmetrical glacial/interglacial cycles) i.e. this is equivalent to saying that average conditions at the sample site have been even colder than the present day upper slopes of Mt Egmont. However, when it is considered that even a 1% error in activation energy will cause a 3°C change in paleotemperature estimate, the estimate is seen in proper perspective. As Williams & Smith (1977, p.124) have stated:

"little credence can be given to absolute average temperatures determined in this manner where accuracies within a few °C are expected".
To illustrate this point further, it might be assumed that the two C\textsuperscript{14} dated Holocene samples (BJP-037 and 040) for which we have D/L ratios (see Table 3.3) would have experienced effective diagenetic temperatures similar to the present mean annual air temperature at the sample site (taken to be c.286°K or 13°C - see Fig.2.8). Therefore calculation of paleotemperatures for these 2 samples should yield temperatures similar to the present. However, because the younger sample (BJP-040) has a higher D/L ratio than the older, let us assume that we have a single sample 7000 yrs old with a D/L ratio of 0.016 (close to the average of the two samples BJP-037, 040). Substitution in equation (2) yields an in situ rate constant of \(K_{\text{L}} = 3.91 \times 10^{-7}\text{yr}^{-1}\), which when substituted in the Arrhenius equation (3) yields an effective diagenetic temperature of 280°K ie. 2°C higher than the 400Ka average derived above.

Interpretation of the above data is as follows:

1. Although the Arrhenius equation yields absolute temperatures which appear c.6°C too low, the calculation of temperature differences is not subject to such large errors.

2. On this basis Holocene ground temperatures have been c.2°C higher than the long term average.

3. To calculate the Holocene time averaged/long term average temperature difference (rather than the rate averaged difference), assume that for simplicity that temperature as a function of time was

\[T_t = T_p\text{ for } t < 10,000\text{ yr.}\]

\[T_p - \Delta T\text{ for } t > 10,000\text{ yr.}\]

McCullough & Smith (1976) have shown that \(\Delta T\) in this situation is given by

\[\Delta T = -RT_p^2 \ln \frac{t_2(K_2/K_1) - 10000}{E} \frac{t_2 - 10,000}{t_2} \]

\]
where $T_p$ is present temperature

- $t_2$ is the age of the pre-Holocene sample (in this case 400Ka)
- $K_1$ is Holocene rate constant = $3.91 \times 10^{-7}\text{yr}^{-1}$
- $K_2$ is long term average rate constant
  \[ = 2.59 \times 10^{-7}\text{yr}^{-1} \]

which yields

\[ \Delta T = 2.3^\circ\text{C}. \]

This value suggests in turn an interglacial-glacial temperature amplitude of c.4.9$^\circ\text{C}$ (Fig. 3.12B), which compares quite favourably with other estimates of this kind.

However, despite the apparent "sensibleness" of the above figure, we have already seen that the modelled Holocene effective diagenetic temperature is c.6$^\circ\text{C}$ lower than expected. This suggests modification of the Arrhenius parameters as follows:

1. Assume the Holocene in situ rate constant is correct, but actually corresponds to a temperature of 286$^\circ\text{K}$ (13$^\circ\text{C}$).

2. The Arrhenius plot for isoleucine kinetics in wood (see Fig. 3.4) after Lee et al. (1976) was based on 3(!) points (see original figure in Lee et al. (1976, p.185, Fig.1)).

3. A simple linear regression through the three data points of Lee et al. (1976) and the Holocene point in Taranaki yields an equation of the form

\[ \log K_{\text{isoleucine}} = 13.0 - 6640(1/T) \quad (R^2 = 1.0) \]

from which

\[ E = 6640 \times 2.303R \]

\[ = 30,400 \text{ cal mol}^{-1} \]

= Arrhenius activation energy

is calculated (cf. 29400 cal mol$^{-1}$ reported by Lee et al. 1976).

Taken as whole the paleotemperature data suggest the dating curve as used here is a satisfactory approximation to the known constraints. More accurate kinetic data would allow further refinement of the curve, and also allow more meaningful paleotemperature estimates.
CHAPTER 4

STRATIGRAPHY OF THE MARINE TERRACES

4.1 PREVIOUS WORK

As part of a regional geological survey, Fleming (1953) described and mapped the distribution of two well developed marine terraces in the Wanganui area: Rapanui (younger) and Brunswick (older). The cover beds of the Rapanui Terrace were designated the Rapanui Formation (Fleming 1946, 1953) and included five named members:

- Kaiwhara Alluvium
- Rapanui Dunesand
- Rapanui Lignite
- Waipuna Delta Conglomerate
- Rapanui Marine Sand

The cover beds of the Brunswick Terrace were named the Brunswick Formation and included six named members:

- Egmont Ash
- Brunswick Dunesand
- Fordell Ash
- Brunswick Beach Sand
- Brunswick Alluvium
- Brunswick Pebbly Sand

For all terrace remnants older than the Brunswick Terrace, Fleming (1953) used the name Kaiatea Terraces. Their cover beds were named the Kaiatea Formation and included one named member:

- Kaiatea Alluvium

Fleming could find no clear evidence of a marine origin for the Kaiatea Terraces and he ascribed to them a fluvial origin. Later work by Grant-Taylor (1964a,b) apparently established their marine origin.
Grant-Taylor also subdivided the Kaiatea terraces into three groups, and these were named from youngest to oldest: Kaiatea III, Kaiatea II, Kaiatea I. No type sections were designated and no map of their distribution was published.

The distribution of the Rapanui, Brunswick and Kaiatea (not subdivided) terraces was shown on the Wanganui 1:250,000 geological map sheet (Lensen 1959).

The term "sub-Rapanui terrace" was used by Fleming (1953) for topographic surfaces that seemed lower than that of the main Rapanui Terrace. This terrace was later redefined by Dickson et al. (1974) as the Rapanui Terrace, and the upper surface was renamed the Ngarino Terrace*. The cover beds of the Ngarino Terrace were designated the Ngarino Formation and included one named member:

Sherwood Sand.

Thus in 1975 the named sequence of terraces (from youngest to oldest) was as follows:

Rapanui
Ngarino
Brunswick
Kaiatea III
Kaiatea II
Kaiatea I

Fig. 8.24 in Suggate et al. (1978) indicated the inferred distribution of shorelines corresponding to each of the above named terraces. In addition, more recent work (McGlone et al. 1979, Aitken in press) has demonstrated the existence of a terrace younger than the Rapanui Terrace. This younger terrace has been informally referred

* Although the Ngarino Terrace was shown to exist as early as 1964 (mentioned by Grant-Taylor (1964a,b), and its inferred distribution in North Taranaki shown on the Taranaki 1:250,000 geological map sheet by Hay (1967), the name was not formalised until 1974 (Dickson et al. 1974).
to as the Inaha Terrace. A marine origin for this, and the other six named terraces has now been generally accepted by all workers, and the terrace sequence is seen as the product of successive relative high sea level events superimposed on a steadily rising landmass.

Two poorly defined biostratigraphic units of the Upper Pleistocene in New Zealand (the Terangian and Oturian Stages) were proposed by Fleming (1953). These were based on faunas within marine deposits at type sections designated on the Brunswick and Rapanui Terraces respectively. The Rapanui and Ngarino Terraces have generally been considered (Fleming 1959, 1975, Grant-Taylor 1964a,b, Suggate 1965,) to have been cut during the last interglacial and the Brunswick Terrace during the penultimate interglacial, although such correlations appear to have been based more on a "count-back-from-present" basis, than on any firm stratigraphic correlations. Dickson et al. (1974) and Suggate (1965, 1978) have entertained the possibility that the Ngarino and Rapanui Terraces might belong to separate interglacials; however, as Suggate (1965, p.81) pointed out, there was a lack of evidence of conditions cold enough to indicate a glacial period between then. On this basis, and accepting correlation of the Terangian and Oturian Stages with the penultimate and last interglacials respectively, Suggate (1965) attempted to redefine the Oturian and Terangian Stages in time stratigraphic terms (see further discussion p.143).

On the basis of inferred stratigraphic relationships between the Brunswick Terrace east of Wanganui and the Mt Curl Tephra (fission track dated at c.230Ka BP), Milne (1973a) suggested that the Brunswick Terrace was cut by an interglacial high sea level between 190 and 230Ka BP and that the Ngarino Terrace stopped forming at or more recently than 115Ka BP. Comparison of these ages with age estimates of glacial terminations in deep sea cores (cf. Broecker & Van Donk
1970) as well as dated sequences of uplifted coral terraces in New Guinea (Bloom et al. 1974, Veeh & Chappell 1970) and Barbados (Mesolella et al. 1969) seemingly supported correlation of the Brunswick and Rapanui Terraces with the penultimate and last interglacials respectively.

The several terrace formations described above have been traditionally assigned to the Hawera Series, which is apparently a biostratigraphic term to cover all beds between the underlying Castlecliffian Stage and the overlying Recent Stage* (Finlay & Marwick 1947). Only the Rapanui Formation is definitely known to overlie the Landguard Formation, the youngest unit of the Castlecliffian Stage, and firm stratigraphic relationships between the top of the Castlecliffian and the other terrace formations have never been established. In fact (Fleming (1953) suggested that) some of the older Kaiatea terraces might be the inland equivalent of the youngest Castlecliffian marine sediments in the Wanganui Basin. More recently, Grant-Taylor & Te Punga (1978 p.552) have reaffirmed that relationships between the Wanganui Series (of which the Castlecliffian is the youngest stage) and the overlying Hawera Series "are not fully resolved".

Table 4.1 summarises existing stratigraphic nomenclature pertaining to the South Taranaki marine terrace sequence.

4.2 STRATIGRAPHIC NOMENCLATURE IN SOUTH TARANAKI

The cover beds overlying the wave cut surface of a particular terrace were seen by previous workers as an easily mappable stratigraphic unit. The cover beds of a terrace were given formation status, and the formation was given the name of the terrace. This was formalised for the Rapanui, Ngarino, Brunswick and Kaiatea Formations;

* Suggate (1962) has advocated inclusion of deposits previously referred to as Recent (Holocene) within the Hawera Series and this is followed here.
however, the subdivision Kaiatea I, II, III (Grant-Taylor 1964a,b) has never been formalised as Kaiatea I, II, III Formations.

Unfortunately the various "formations" of terrace cover beds do not strictly qualify as lithostratigraphic units for the following reasons:

1. Terrace morphology alone may be used for their identification and correlation away from type localities. They are therefore best viewed as morphostratigraphic units, and should not bear the name "formation". (According to Bowen (1978, p.94), "morphostratigraphic units are of informal status only...and their terminology should not be applied with any stratigraphic implication").

2. The cover beds of a particular terrace lack any unifying lithological characteristic(s), whereas according to Hedberg (1976, p.31) the critical requirement of a lithostratigraphic unit is a "substantial degree of overall lithologic homogeneity".

Consequently it is here proposed to redefine the "lithostratigraphic units" in more rigorous terms.

On the other hand, Hedberg (1976, p.43) states that previously named stratigraphic units of wide acceptance should be retained wherever possible, even if their naming does not strictly conform to latest stratigraphic nomenclature.

In the light of the above considerations, previous stratigraphic names will not be abandoned and redefinition will be as follows: The Brunswick Formation, for example, which previously included all deposits (both marine and non-marine) overlying the Brunswick wave cut surface, will be redefined to apply only to the marine sediments within the Brunswick Terrace. Several new formations will be proposed for the marine cover beds of newly named terraces. These formations will also bear the name of the terrace eg. Inaha Formation will be proposed for marine cover beds overlying the wave cut surface of the Inaha Terrace. Such redefinition means that formations would be restricted to individual marine terraces, whereas previously non-marine fluvial, laharic and tephric beds within formations
could cross terrace boundaries. Furthermore, although Hedberg (1976, p.31) has pointed out that lithostratigraphic units should not be recognised and defined by inferred geologic history or mode of genesis; it is emphasised here that the marine sands within terraces in South Taranaki may be adequately defined in terms of their observable physical features (see 4.3.2 Sedimentary environmental diagnostics, and Table 4.2).

4.3 FIELD MAPPING

4.3.1. METHODS

Field work concerned with terrace stratigraphy was carried out in two major ways:

1. Description and correlation of sections exposed in road cuttings, railway cuttings, farm tracks, soil slips and other natural exposures such as coastal cliffs or river valleys. Most sections were sketched (generalised in some cases) in field notebooks and both thickness and lithology of individual units recorded. Thicknesses were usually estimated visually although important sections were logged with tape measure, or in the case of thick sections, with a Paulin surveying altimeter (limit of reading ±1m).

2. Identification of wave cut surfaces, accompanied by height estimation using surveying altimeter. Surveying was tied to trig points or bench marks where possible, although in many instances spot heights at road intersections, given on NZMSI 1:63,360 series topographic maps, were used (accuracy of spot heights ±1.5m normally). Allowance was made for air pressure variations by assuming linear drift between points of known
height. Since survey traverses were kept short, accuracy of surveyed heights is thought to be better than ±3m.

Altimetric survey of wave cut surfaces and sections was important because:

a) Considerable differential uplift has occurred parallel to the present coastline, and simple altimetric correlation of terraces is not possible. Faults also offset the terrace surfaces.

b) Heights of terrace surfaces are not necessarily a good indication of wave cut surface or strandline heights.

c) Fossil cliffs are rarely exposed; however, their presence may be inferred by abrupt height differences in wave cut surfaces.

Map 1 shows the extent of each mapped terrace, including all surveyed wave cut surface heights. Type sections and proposed type sections are also shown, as well as faults and fossil cliffs.

4.3.2. SEDIMENTARY ENVIRONMENTAL DIAGNOSTICS

As indicated previously, a wide variety of sedimentary units overlies the South Taranaki terraces. Here I describe some of the ways of interpreting their depositional environments. In particular, the distinction between aeolian, beach, fluvial, estuarine and shallow marine sandy sediments is important because these are widespread and yet easily confused.

As Gow (1967) has shown, the principal source of beach sediment near Hawera is that derived from weathering and erosion of volcanic deposits on the Taranaki Peninsula. These andesitic materials are brought to the coast by rivers draining the slopes of Egmont, Pouakai and Kaitake and are carried along the coast in the nearshore zone to produce the characteristic black sand beaches in Taranaki. The principal dark minerals present in the beach sands are titanomagnetite, augite and hornblende (with lesser amounts of hypersthene and olivine), the percentage of which decreases north and south with increasing distance from source. Fleming (1953 p.285) reported that
beach sand at Nukumaru contained 65% magnetite grains, from Kai-Iwi Beach 31.8% and from 4 km north west of Wanganui, 8.5% magnetite. According to Gow (1967) erosion and reworking of the cover beds of the Rapanui Terrace is also an important source of andesitic minerals in the present day beaches near Hawera.

Contrasting with the andesitic beach sands and river sediments of the western Taranaki area are fluvial sediments derived from the denudation of the Cainozoic shallow marine sediments of the Wanganui Basin. These are typically quartzose-feldspathic sediments with prominent mica, and virtually no dark minerals. Some hypersthene is derived from central volcanic region and is brought down rivers east of and including the Wanganui (Fleming (1953)).

In general, therefore, between Hawera and Wanganui two major classes of sediments can be distinguished in terms of provenance, as follows:

(a) Ferromagnesian rich eg. beach and coastal dunes, shallow marine.
(b) Ferromagnesian poor eg. fluvial and estuarine, while the provenance of offshore marine remains uncertain.

There is a vast literature concerning the reconstruction of sedimentary environments on the basis of sediment characteristics such as grain size, sedimentary structures and lithology. Of major interest here, however, are characteristics which aid in the field interpretation of sedimentary environments - in particular sedimentary structures and lithology. Table 4.2 presents this information in summary form. While no single feature is necessarily diagnostic, several features taken together allow a reliable interpretation to be made. The list is not exhaustive, and further work will undoubtedly add to and modify the contents.

Some useful references include McKee (1957), Glennie (1970), Reineck & Singh (1973), Conybeare & Crook (1968), Rigby & Hamblin
(1972), Christie (1979); however, many of the diagnostics are of local use only, and are not applicable necessarily to similar depositional environments elsewhere.

4.4 TEPHROSTRATIGRAPHY

4.4.1 TEPHRAS AND THEIR USES

"Tephra" is a term originally coined by Thorarinsson for "all the clastic volcanic material which during an eruption is transported from the crater through the air, corresponding to the term lava to signify all the molten material flowing from the crater" (Cole & Kohn 1972 p.686). As pointed out by Healy (1972) it is a particularly useful term since it avoids the inconsistent and loose use of the word "ash" to describe deposits that, near source, include ejectamenta of all size classes. The term is applied not only to airfall pyroclastic deposits (tephra falls) but also to pyroclastic flow deposits which are the product of nuees ardentes (tephra flows).

Some of the most valuable chronologic and stratigraphic markers in New Zealand Quaternary studies are widely distributed tephras in the North Island. This is because they may be considered as virtually instantaneous geological events. Furthermore they enable detailed correlation over wide areas and once a tephra has been dated at one site, it can then be used as a time plane at other sites. On the other hand there are difficulties such as the change in character of a tephra with increasing distance from source as well as the effects of differing weathering intensities. Also many tephras are superficially similar, particularly when weathered.
Important tephras in the North Island are generally of two major types: rhyolitic or andesitic. Rhyolitic tephras erupted from sources within the Taupo Volcanic Zone tend to produce much more voluminous and widespread deposits than andesitic tephras. Some of the rhyolitic tephras can be detected as macroscopic layers in deep sea cores hundreds of kilometres east of New Zealand (Ninkovitch 1968, Watkins & Huang 1977) and were almost certainly associated with ignimbrite eruptions. Andesitic tephras, on the other hand, are more valuable on a local scale eg. around Mt Egmont (Neall 1972) and Mt Tongariro (Topping 1973).

A large number of Taupo Volcanic Zone rhyolitic tephras have been identified and dated within the last 40,000 years using C\(^{14}\) dating methods on organic material from within or immediately underlying individual tephras (eg. Vucetich & Pullar 1969). While many of the larger eruptions within this period produced tephras which have been recognised and traced over large areas in the central and eastern parts of the island, only one (the Kawakawa Tephra Formation dated at circa 20,000 years BP) is macroscopically observed in South Taranaki. This has arisen largely as a result of the predominantly west to east atmospheric circulation over New Zealand which tends to carry volcanic airfall material eastwards. Much more important in South Taranaki (particularly in the western part), have been Egmont-source andesitic tephras, partly owing to proximity and partly because of the downwind location of the terraces.

In this study, only one prominent rhyolitic tephra older than the Kawakawa has been observed to occur within the South Taranaki terrace sequence. This tephra, named the Omahina Tephra in this work, occurs within the cover beds of older terraces of the sequence, but is absent from the younger ones. Fresh volcanic glass and zircons were
extracted from the Omahina Tephra at one site, and these were dated by the fission track method. The Omahina Tephra may be distinguished from the Kawakawa by its coarser grainsize and much lower stratigraphic position within terrace cover beds. Andesitic tephras, while present in large numbers within the terrace sequence, were not so useful in this study because they:
   (a) are all quite similar in appearance
   (b) are generally weathered
   (c) by their very nature, are not suitable for fission track dating
   (d) thin very rapidly away from source.
However, future work will probably show them to be valuable marker beds at least in the western half of the terrace sequence.

4.4.2 OMAHINA TEPHRA (new name)

4.4.2.1 Type Section:
The type section is a road cutting on Omahina Road at grid reference N130/237150 (Plate 4.2, Fig. 4.1).

4.4.2.2 Name:
After Omahina Road on which the type section occurs.

4.4.2.3 Reference Sections:
N129/948367 Morea Road
N129/013299 Meremere
N129/087238 Ball Road Extension
N130/122178 Opaku (Elliot Road)
N130/184160 Kohi Road
N130/296120 Ridge Road
N137/471066 Rangitatau East Road
N119/904423 Rotokare Road

4.4.2.4 Nature and Distribution
The known distribution of the Omahina Tephra is shown in Fig. 4.1, together with five representative stratigraphic sections. The tephra has been identified at 15 localities (see Fig. 4.1, Plates 4.1 and 4.2) as well as Appendix 2 for details). Identification is least
certain at localities 1 and 14. Field identification and correlation is aided by the presence of a distinctive, dark brown, weathered andesitic ash which overlies the Omahina Tephra at many localities. The pale colour of the Omahina Tephra, and the presence of its characteristic coarse, sandy basal layer (dominantly plagioclase feldspar grains) are also useful distinguishing features.

At all localities, except Locality 1 where it has not yet been demonstrated, the Omahina Tephra underlies terrace surfaces older than those of the Brunswick Terraces. Although many well exposed sections have been examined, the tephra has not been recorded within the cover beds of either of the Brunswick Terraces. The youngest marine terrace within which the tephra occurs, is the Kaiatea III Terrace.

At the type section (Fig. 4.1, Plate 4.2) the Omahina Tephra is a 15cm thick subhorizontal layer. It is a single graded bed, containing fragments up to 2mm in diameter in its coarsest basal part. At this, and other sections, the upper, finer parts of the tephra are weathered to a sticky cream coloured clay which has been identified as halloysite (V. Neall pers. comm. 1977). Mineralogical analysis shows that the Omahina Tephra, at its type section, contains abundant rhyolitic glass, common plagioclase, secondary hypersthene, scarce quartz, titanomagnetite, hornblende and rare zircon and biotite. This mineral assemblage indicates that the tephra is the product of a rhyolitic volcanic eruption, probably the airfall equivalent of one of the central North Island ignimbrite sheets.

* A small terrace remnant at the back of the Brunswick Terrace is intermittently present between Waverley and Manganui. To emphasise its close association with the Brunswick Terrace, it will be named in a later section (see page 88) as the Upper Brunswick Terrace. Collectively the two terraces will be referred to as "the Brunswick Terraces" (see Fig. 4.1 and Map 1 for distribution).
At the type section, the Omahina Tephra closely overlies well sorted andesitic sands within the Kaiatea III Terrace cover beds. A wave cut surface overlain by pebbly sands is poorly exposed in a small stream valley some 200m south (NI30/237147) of the type section, and lies some 16m below the Omahina Tephra. The well sorted andesitic sands (grey) are interpreted as the upper part of neashore marine or littoral sediments conformably overlying the Kaiatea III wave cut surface. A similar, but more completely exposed section occurs on Ridge Road some 6km to the east (Locality 12, Fig. 4.1), where a thick sequence (14m) of laminated, pebbly marine sands overlies a bored siltstone surface and is closely overlain by the Omahina Tephra. At both the above localities, the Omahina Tephra lies less than 2m above the top of the marine sediments and cutting of the Kaiatea III wave cut surface is therefore inferred to closely precede the fall of the Omahina Tephra.

The tephra has been tentatively identified at one locality in North Taranaki (Locality 1, Fig. 4.1), where it occurs within a marine terrace inland from Urenui. Further work in this area should ultimately permit correlation from South to North Taranaki using the Omahina Tephra as a stratigraphic marker.

4.4.2.5 Age:

Non-pumiceous glass shards and zircon crystals (63-250 microns) were extracted from a bulk sample of the basal layer of the Omahina Tephra at its type locality using conventional heavy liquid and magnetic separation techniques. Samples were kindly fission track dated by Dr B.P. Kohn, Ben Gurion University of the Negev, Israel. Dating methods are described by Pillans & Kohn (in press). Table 4.2 shows fission track data and dates reported by Pillans & Kohn for the Omahina Tephra, Rangitawa Pumice and Mt Curl Tephra. Errors on all
dates are expressed at ±2σ and were calculated on the number of spontaneous and induced tracks counted, and the number of tracks counted in the standards.

Concordant fission track ages were obtained on both zircon (370 ± 50 ka) and glass (370 ± 70 ka) from the Omahina Tephra.

4.4.2.6 Correlation:

Pillans & Kohn (in press) have correlated the Omahina Tephra with the Rangitawa Pumice on the basis of:

1. the similarity in fission track ages at Omahina Road (type locality of Omahina Tephra) and those from Rangitawa Stream (type locality of the Rangitawa Pumice), as well as the marked discordance with dates on the Mt Curl Tephra from its type locality (see Table 4.2 for summary of dates),

2. the similar ferromagnesian mineralogy of the tephras at Omahina Road and Rangitawa Stream, both of which are dominantly hypersthene with secondary hornblende.

Pillans and Kohn suggest that further dating, mineralogy and chemistry of glass and ferromagnesian minerals are required to more fully test the proposal.

The Rangitawa Pumice has previously been only positively identified at its type locality (N143/962624) in the Rangitikei Valley (Fig. 4.1), although Pillans & Kohn (in press) report tentative identification at one other locality (N137/762873), as well as the 15 localities at which the Omahina Tephra has been identified. The only published description of the type locality of the Rangitawa Pumice is a generalised one given by Te Punga (1962) as follows:

Halcombe Conglomerates unconformably overlying
20' Rangitawa Pumice (white rhyolitic)
15' yellow-brown stratified sand with lenses of small pebbles
10' Rangitawa Fossil Beds; richly fossiliferous, blue-grey sandy mudstone with some yellowish muddy sandstone.

More recently Seward (1976) has reported a thickness of 1.5m for the Rangitawa Pumice at its type locality (see Plate 4.3).

At its type section the Rangitawa Pumice has a measured thickness of 70cm and is directly overlain by 80cm of pale weathered silts containing rootlet impressions. These silts are overlain by laminated pebbly sands interpreted to be marine*, and which are in turn overlain and truncated by a coarse fluvial conglomerate. The Rangitawa Pumice therefore appears to be conformably underlain and overlain by Upper Castlecliffian marine sediments (see also Te Punga 1962 and Boellstorff & Te Punga 1977). The Rangitawa Fossil Beds described by Te Punga (1962) appear to be no longer exposed (well preserved mollusc valves are present in the bed of Rangitawa Stream for several hundred metres upstream from the type section, suggesting that the fossil beds might be exposed further upstream).

The inferred stratigraphic position of the Mt Curl Tephra† east of Wanganui as described by Milne (1973) is identical to that of the Omahina Tephra in as much as both tephras appear to occur within marine terraces older than the Brunswick Terrace. However, as indicated by Pillans & Kohn (in press) (see also Table 4.2), fission track ages on both the Mt Curl and Omahina Tephras are statistically quite different and do not support their correlation: samples of both tephras were irradiated together so that any apparent age differences arising from sample preparation could be minimised (BP Kohn pers.comm. 1979). On the other hand Seward (1976) has dated a rhyolitic ash, called the Lower Finnis Road Ash from a locality some 30km east of

* The continuity and similarity of sediments upwards from the lower part of the exposed section, just below which marine fossils reported by Te Punga (1952, 1962), suggests a marine origin. The lateral extent of laminations, and lack of scour and fill structures support this interpretation.

† See Appendix 3
Rangitawa Stream which gives a fission track age on glass of $320^{+70}_{-10} \text{Ka}$. Statistically the dates for both the Lower Finnis Road Ash and the Omahina Tephra are indistinguishable, and in view of the hypersthene/hornblende dominated ferromagnesian assemblage for both ashes, their correlation cannot be ruled out. (Like the Rangitawa Pumice, the Lower Finnis Road Ash occurs within younger Castlecliffian shallow marine sediments of the Wanganui Basin).

4.5 TERRACE STRATIGRAPHY

The stratigraphy of individual terraces is presented in this section, in chronological order from oldest to youngest.

4.5.1 PRE-KAIATEA TERRACES

Dissected remnants of terrace surfaces which predate the Kaiatea I Terrace are locally present, particularly east of the Nukumaru Fault Zone (Fig. 4.2, Map 1). At least two older terraces appear to be present, although exposures are limited and prevent more certain conclusions from being drawn.

An erosional unconformity poorly exposed near Marorau Road (N138/568042) is interpreted to represent a wave cut surface underlying the youngest of these pre-Kaiatea terraces, and occurs at an altitude of 309m. A steep topographic rise trending east-west occurs some 500m north of this exposure and is thought to represent the surface expression of the terrace cliff and strandline.

On Rangitatau East Road at grid reference N137/476074, pebbly sands overlie an erosional unconformity at an altitude of 364m. Once again, immediately inland there is an abrupt topographic rise, which is interpreted to be the surface expression of a fossil cliff.
Approximately 1km south of this exposure, the Omahina Tephra is exposed at grid reference N137/471066 at an altitude of 359m. Thick sequences of weathered andesitic ashes are exposed in road cuttings here and to the south; however, no wave cut surfaces are exposed and this prevents assignment of these cuttings to either of the above mentioned terraces.

Extrapolated strandline height estimates suggest that the type locality of the Kaiatea Formation (terrace remnants near the Kaiatea Trig Station of grid reference N138/643007) is intermediate in altitude between the Kaiatea I marine Terrace and the older unnamed remnants. The type exposure of the Kaiatea Alluvium (named member of the Kaiatea Formation) also occurs there. This locality, which occurs on a river terrace adjacent to the Wanganui River, was designated type locality for all terraces older than Brunswick by Fleming (1953) who could find no evidence of their marine origin. New type localities will be designated for the Kaiatea I, II and III terraces in later sections.

4.5.2 KAIATEA I TERRACE AND FORMATION (new name)

4.5.2.1 Type section:
The type section is a road cutting on Okahutiria Road at grid reference N130/187187.

4.5.2.2 Name:
After Kaiatea Trig. Station at grid reference (N138/652007) in the Wanganui River Valley, where Fleming (1953) originally designated the type locality of the Kaiatea Formation. The Kaiatea
terraces were later subdivided as Kaiatea I, II, III by Grant-Taylor (1964a,b), but no type localities were given.

4.5.2.3 Reference Sections:

N137/466044 Rangitatau East Road
N130/087238 Ball Road Extension (Plate 4.1)

4.5.2.4 Nature and Distribution:

The Kaiatea I Formation is here proposed to include all marine sediments overlying the Kaiatea I wave cut surface. Its lower boundary is therefore an erosional unconformity. At the type section this erosional unconformity is overlain by quartzite pebbles at an elevation of 330m. Immediately overlying these are gravelly sands (80cm thick) containing distinctive zoned siltstone clasts (Fig. 4.2 and Plate 4.4). The zoning appears to be a weathering feature. A total thickness of 3m of laminated pale marine sediments overlies the erosional unconformity (Fig. 4.2), and these are in turn overlain by weathered dune sand and a thick sequence of weathered ashes. Small gully fills are evident in the upper portions of the type section, suggesting that the section has been truncated in the past. At this and other sections the concentration of mafic minerals in the marine sands is low.

The Kaiatea I Terrace, is here named to include all deposits and features overlying and including the Kaiatea I wave cut surface. The terrace has been extensively dissected by streams, and consists largely of scattered hilltop remnants at many localities. It occurs immediately inland of the Kaiatea II Terrace (where present) or the Kaiatea III Terrace where the Kaiatea II Terrace is absent (removed by subsequent erosion during the cutting of the Kaiatea III wave cut surface).
The distribution of the Kaiatea I Terrace is shown in Fig. 4.2, surveyed wave cut surface heights are also shown.

4.5.3 KAIATEA II TERRACE AND FORMATION (new name)

The Kaiatea II Marine Terrace is a discontinuous terrace which occurs at the outer edge of the Kaiatea I Terrace (Map 1; Fig. 4.2). Very few good exposures of the wave cut surface and cover beds are known.

4.5.3.1 Type Section:

No suitable section is known. The section exposed in a soil slip near Meremere (N129/013299) is designated as a reference section (Fig. 4.1).

4.5.3.2 Name:

As for Kaiatea I Terrace and Formation (see page 84).

4.5.3.3 Reference Section:

N129/013299 Meremere (Figs. 4.1, 4.2).

4.5.3.4 Nature and distribution:

The Kaiatea II Formation is here named to include all marine sediments overlying the Kaiatea II wave cut surface. Lithology and general character appear similar to the Kaiatea I Formation.

The Kaiatea II Terrace is here defined to include all deposits and features overlying and including the Kaiatea II wave cut surface. The c.4m thick lignite exposed at the reference section, is informally referred to as the Meremere Lignite (Fig. 4.1).

Surveyed wave cut surface heights are shown in Fig. 4.2.
4.5.4 KAIATEA III TERRACE AND FORMATION (new name)

4.5.4.1 Type Section:
The type section is a road cutting on Morea Road at and near grid reference N129/948367 (Figs. 4.1, 4.2).

4.5.4.2 Name:
As for Kaiatea I Terrace and Formation.

4.5.4.3 Reference Sections:
N129/013259 Ingahape Road
N130/122178 Opaku (Elliot Road)
N130/184160 Kohi Road
N130/296120 Ridge Road
N130/237150 Omahina Road
N137/478014 Rangitatau East Road

4.5.4.4 Nature and distribution:
The Kaiatea III Formation is here named to include all marine sediments overlying the Kaiatea III wave cut surface. The lower boundary is therefore an erosional unconformity. Characteristically, the formation consists of laminated micaceous sands with rare scattered pebbles. A thin basal conglomerate is usually present. No marine fossils have been recorded, although the underlying rocks are locally bored eg. Ridge Road Section N130/296120 (Fig. 4.1). The formation has a maximum inferred thickness of 15m at Omahina Road (N130/237147) and a maximum exposed thickness of 12m at Ridge Road (N130/296120) - see Figs. 4.1, 4.2). The formation may be distinguished from underlying strata by the erosional unconformity (if exposed) and by an increased abundance of mafic minerals, particularly near the base. The content of mafic minerals is higher than either
the Kaiatea I or II formations (Plate 2.12) and increases towards the western end of the terrace sequence.

The Kaiatea III Terrace is here defined to include all deposits and features overlying and including the Kaiatea III wave cut surface. Above the Kaiatea III Formation, this includes a variable sequence of terrestrial deposits (eg. dunesand, weathered andesitic ash, fluvial sediments), within the lower part of which occurs the Omahina Tephra. At the type section (Figs. 4.1, 4.2) on Morea Road the thickness of terrestrial sediments overlying the Kaiatea III Formation exceeds 45m, and total cover bed thickness is 68m. Approximately 50% of this thickness is not exposed. The thickness of cover beds within the terrace decreases eastwards from the type section, as the input of material from the Taranaki volcanoes decreases further from source. Average estimated cover bed thickness inland from Waverley is c.30-35m, or about half that at the type section.

The distribution of the Kaiatea III Terrace is shown in Fig. 4.2. Representative stratigraphic sections are shown in Figs. 4.1 and 4.2. Surveyed wave cut surface heights are also shown in Fig. 4.2.

4.5.5 UPPER BRUNSWICK TERRACE AND FORMATION (new name)
The Upper Brunswick Terrace is named here for a small terrace remnant which occurs intermittently at the inland margin of the Brunswick Terrace (see Fig. 4.3 and Map 1). It reaches a maximum width of 700m, inland from Waverley, near grid reference N130/210150 (cf. 2000m width of nearby Brunswick Terrace).
4.5.5.1 Type Section:
Lack of good exposure, and the limited extent of the terrace prevent designation of a type section at this time. A farm track leading from Braemore Road is named as a reference section (N130/211150).

4.5.5.2 Name:
To emphasise the close association with the previously named Brunswick Terrace (Fleming 1946, 1953), the prefix "upper" is added which is indicative of its higher elevation in relation to the Brunswick Terrace.

4.5.5.3 Reference Section:
N130/211150 Braemore Road

4.5.5.4 Nature and description:
The Upper Brunswick Formation is named here to include all marine sediments overlying the wave cut surface within the Brunswick Terrace. Its lower boundary, which is an erosional unconformity, is exposed at the reference section (N130/211150), where 3m of laminated sands (pebbly in the basal 60cm) overlie a wave cut surface in yellow micaceous sands at an elevation of c.200m. A similar exposure occurs on the western side of Huritoia Stream (N137/462010), where 3m of laminated, pebbly, andesitic sand overlies a wave cut surface in shell-rock at an altitude of 164m. A thin (2cm) weathered white ash overlies these marine sands at this locality but correlation is not attempted at this stage. Further east on Koatanui Road a total thickness of 26m of cover beds has been measured for the Upper Brunswick Formation (N137/512006). At the reference section total cover bed thickness is c.23m. These figures suggest that the average cover bed thickness is similar to that within the Brunswick Terrace.
Despite the lack of good exposures, and the limited extent of the terrace, it is nevertheless quite a prominent feature when present (Plate 4.6). Immediately west of Omahina Road near grid reference N130/237147, a prominent cliff of c.20m is cut in Pliocene siltstone and separates the Upper Brunswick from the next oldest terrace, the Kaiatea III Terrace. Here the Upper Brunswick strandline reaches 213m and the wave cut surface at the outer edge of the Kaiatea III Terrace is at 241m. Similarly a prominent, but smaller c13m cliff cut in siltstone is exposed near N130/236147, and marks the Brunswick/Upper Brunswick cliff. Here the Brunswick strandline reaches 193m and the wave cut surface at the outer edge of the Upper Brunswick Terrace is at 206m. The similarity in heights of the Brunswick and Upper Brunswick strandlines here (193 and 213m respectively) and elsewhere, suggest that the two terraces are similar in age. (Similarity in cover bed thicknesses supports this interpretation, although detailed comparison of Brunswick/Upper Brunswick cover beds is not yet possible owing to poor exposure within the Upper Brunswick).

The best topographic expression of the Upper Brunswick Terrace may be seen driving up Braemore Road, where the fossil cliffs at its inner and outer edge are quite marked (grid references N130/209156 and N130/210149 respectively) - see Plate 4.6.

The known distribution of the Upper Brunswick Terrace is shown in Fig. 4.3; surveyed wave cut surface heights are also shown.

4.5.6 THE BRUNSWICK TERRACE AND FORMATION (Fleming 1953, redefinition)

The Brunswick Terrace is the most easily mappable of the South Taranaki Terraces for several reasons:

1. It is truncated at its outer edge by the Ngarino cliff, which is the highest and most prominent on the terrace sequence (Plate 2.9). The Brunswick cliff at the inner edge of the Brunswick Terrace is also quite prominent.
2. It is dissected by streams to such a depth and extent that good exposures of the cover beds and wave cut surface are readily available (Plate 2.8). On the other hand the terrace is not so dissected as to leave only minor remnants.

3. The marine sediments contain substantial amounts of mafic minerals (sands) and andesite pebbles (conglomerates). This makes them readily distinguishable from underlying sediments.

4. The stratigraphy differs from the Kaiatea Terraces in that the Omahina Tephra is not present within the Brunswick cover beds at any locality. Average cover bed thickness is estimated to be c.25m.

4.5.6.1 Type Section:
The type section is a road cutting on the Kai-Iwi to Brunswick Road at grid reference N137/502981 (Fleming 1953).

4.5.6.2 Name:
After the settlement of Brunswick, some 5km east of the type section.

4.5.6.3 Reference Sections:
N129/056215 Makino Road
N129/057177 Ohaka Stream
N129/941345 Tangahoe Valley Road
N130/123158 Opaku
N137/378002 Ototoka Stream
N137/420992 Ototoka Stream
N137/428997 Pukerimu Road
N137/509947 Stannard Road
N138/582940 Brunswick Road Quarry
N138/732815 Warrengate
N138/759838 Kauangaroa Road
N138/771851 Kauangaroa Road

4.5.6.4 Nature and distribution:
Fleming (1946 p.348) proposed the name Brunswick Formation for the sediments forming the cover of the Brunswick Terrace. No type
section was designated; however, it was later proposed (Fleming 1953 p.252) as a road cutting on the Brunswick to Kai-Iwi Road near grid reference N137/502981.

The Brunswick Formation is here redefined to include only the marine sediments overlying the Brunswick wave cut surface. It is therefore equivalent to the lowermost named member of Flemings (1946, 1953) Brunswick Formation, the Brunswick Pebbly Sand.

At the type section, Fleming (1953) described and named five members of the Brunswick Formation (sensu lato) as follows (oldest to youngest) - see Fig. 4.4):

1. BRUNSWICK PEBBLY SAND, the lowermost unit of marine origin. According to Fleming (1953 p.253) the lowest bed of the Brunswick Pebbly Sand consists of "3ft of white, cross-bedded, medium-to coarse-grained, loose sand, with rare conglomeratic layers containing quartzite pebbles, and is overlain by a bed of fragments of pumice siltstone up to 8in long". However, close inspection of the type section (recent road works have left excellent exposure) reveals this unit is part of the underlying Nukumaruan sequence. This assessment is also supported by Flemings (1953, p.253) further observation that: "overlying beds contain ferromagnesian minerals, magnetite and pebbles of augite- hornblende andesite and hypersthene andesite" i.e. that they are mineralogically quite different from underlying sediments.

The lower boundary of the Brunswick Formation is here redefined as the erosional unconformity (wave cut surface at c.129m) which separates the laminated pebbly sands containing
igneous pebbles and mafic minerals, from the underlying loose pumice sands. The upper boundary is defined as the base of the carbonaceous silt bed which overlies low angle cross-bedded sands some 5m above the wave cut surface (see Fig. 4.5 for comparison between present interpretation of the type section and that described by Fleming (1953)). A thin layer containing shell grit some 3m above the base of the formation appears to correspond with a shell grit layer mentioned by Fleming (1953, p.253, Fig. 47 and faunal list p.260-261). The shells are poorly preserved fragments occurring in ferruginised gravelly sands.

The Brunswick Pebbly Sand is here redefined as the Brunswick Formation *sensu stricto*, and it is therefore recommended that further use of the term be discontinued.

2. *BRUNSWICK ALLUVIUM*, consisting of "variable fluviatile, lagoonal or deltaic sediment..." (Fleming 1953, p.253). This unit overlies the Brunswick Formation and may be distinguished from it by the lack of mafic minerals and generally finer grain size. The character of the sediments and the lateral continuity (>20m) of thin laminations suggests a low energy environment of deposition, perhaps an estuary. A 10-30cm thick lignite bed containing wood occurs some 100cm above the base of the unit. Further use of the formal name Brunswick Alluvium is to be abandoned. This and overlying units within the Brunswick Terrace will be informally referred to as Brunswick cover beds until more detailed work enables appropriate formal lithostratigraphic subdivision and naming.
3. BRUNSWICK BEACHSAND, with a "high degree of sorting and concentration of magnetite in its finer beds" (Fleming 1953, p.253). According to Fleming (1953, p.253) this unit is "a fairly persistent member west of the Wanganui River, where it grades up conformably from the Brunswick Pebby Sand". Re-examination of the type section reveals that this unit is the basal portion of overlying dunesands. Parallel lamination and low angle cross beds are present, and are suggestive of a marine origin. However, the well sorted nature of the sand, its lack of pebbles or marine fossils, and its high concentration of mafic minerals are more in keeping with an aeolian origin (see Table 4.2). I therefore propose that usage of the term Brunswick Beach Sand be discontinued.

4. FORDELL ASH, "a light ashy silt, 3 to 5ft thick" (Fleming 1953, p.253) supposedly overlies Brunswick Beach Sand at the type section. The unit was not recognised by me on examination of the section, and its designation remains uncertain. Fleming clearly used the name "Fordell Ash" as a composite term for a variety of sediments which included not only at least two rhyolitic tephras, but also interbedded andesitic tephra, silts and dunesand. He designated sections exposed on No.2 Line Extension (now Kauangaroa Road) 20km east of Wanganui as the type sections. Two prominent rhyolitic tephras are present in sections (N138/732815, N138/759838) on Kauangaroa Road. However, further work is clearly needed to clarify the situation there, and also to test the correlations made by Fleming west of Wanganui. I would expect that the term "Fordell Ash" will eventually be redefined to name one or other of the prominent tephras on Kauangaroa Road.
5. BRUNSWICK DUNESAND, consisting of "impure iron sand of medium grainsize, which displays aeolian bedding in good exposures" (Fleming 1953, p.254). This unit, which is here recommended to be of future informal status, can be redefined to include all dark grey cross-bedded sands above the Brunswick Alluvium at the type section. The unit is overlain by up to 4m of strongly weathered clays which could be tephra, loess or a mixture of both. Two thin interbedded, weathered tephras are present at the type section in the upper part of the Brunswick Dunesand (Fig. 4.4).

According to Fleming (1953, p.253 Fig. 47), the uppermost member of the Brunswick Formation is Egmont Ash, which was the name given to "superficial andesitic ash-beds with augite, hornblende and magnetite prominent among their heavy minerals, that overlie the coastal terraces and old land-surfaces of Wanganui subdivision (Fleming 1953, p.282). By this definition Egmont Ash was also inferred to be a member of both the Rapanui and Kaiatea Formations (sensu lato), although Fleming (1953) stated elsewhere (p.249 and p.263) that it formed a superficial veneer on both these formations. Table 14 in Fleming (1953, p.103) implies that Egmont Ash was not a member of the Brunswick Formation sensu lato. It has therefore remained an obscure stratigraphic term, and I recommend here that its use be completely discontinued. The name "Egmont Ash" should not be confused with the "Egmont Shower" which occupies an important stratigraphic position closer to Mt Egmont (Neall 1972).

In summary, all units named by Fleming (1953) as members of the Brunswick Formation are to be renamed, abolished or have informal
status as follows: marine sediments overlying the Brunswick wave cut surface are redefined as the Brunswick Formation (sensu stricto) - previously Brunswick Pebby Sand. All non-marine coverbeds overlying the Brunswick wave cut surface are informally referred to as Brunswick Coverbeds - previously Brunswick Alluvium (now informal usage), Brunswick Beach Sand (no longer recognised), Fordell Ash (status uncertain), Brunswick Dunesand (now informal usage) and Egmont Ash (name to be discontinued).

East of Wanganui the Brunswick Terrace was mapped and described by Fleming (1953), and he described a particularly well exposed section through the coverbeds near Warrengate Trig Station at grid reference N138/732815 (see Fleming 1953: description p.258, which appears incomplete, and text figure 48, p.257 do not correspond exactly). The log made by Fleming (1953) is redrawn in Figure 4.5 together with a log made during this study. The section is important because it demonstrates stratigraphic relationships between the Brunswick Terrace and uppermost marine sediments of the Wanganui Basin.

An erosional unconformity at 114m in the Warrengate Section is cut in lithified, laminated silts and is directly overlain by a thin layer of carbonaceous silt and black pebbly sands (Fig. 4.5). These lithified silts are unlike any of the sediments within the Brunswick Terrace deposits at this and other sections, but the degree of lithification is similar to Plio-Pleistocene siltstones underlying the terrace at many other localities. This erosional unconformity, which is locally bored, is interpreted to be the wave cut surface at the base of the Brunswick Formation. The thickness of cover beds overlying this wave cut surface (27m) is very similar to cover bed thickness northeast of the Warrengate Section (illustrated by Fleming
1953 Fig. 48 p.257). Some 5m above the wave cut surface, a prominent, weathered silt with root cavities is interpreted to be a fossil soil, and is taken to represent the top of the marine sediments overlying the cut surface. The Brunswick Formation at this section is therefore c.5m thick.

Some 14m below the base of the Brunswick Formation at the Warrengate Section, a prominent shellbed is exposed which contains the marker fossil *Pecten novozelandae aotea*. This pecten is an index fossil of the Landguard Formation, the youngest formation of the Castlecliffian Stage. A further 12m below this shellbed is another shellbed containing *Pecten benedictus marwicki*, which has also been reported by Fleming (1953, 1957) from the Putiki Shellbed and the Upper Castlecliff Shellbed near Wanganui. The Putiki Shellbed closely underlies the Landguard Formation in the standard Castlecliffian sequence near Wanganui (see Table 4.1), and is the most likely correlative of the lower shellbed at Warrengate. Correlation of the upper shellbed at Warrengate with the Landguard Formation, means that for the first time the Brunswick Formation can be unquestionably demonstrated to be younger than the top of the Castlecliffian Stage.

The Warrengate Section is not well exposed between the interpreted base of the Brunswick Formation and the shellbed containing *P.n. aotea*; however, the limited exposure allows the following conclusions to be drawn:

1. Unlike sections to the west of Wanganui, there is considerable andesitic-source sediment below the Brunswick wave cut surface.
2. Another wave cut surface (bored siltstone) which lies some 3m above the shellbed records another regression/transgression cycle between the time the shellbed was deposited and the time the Brunswick Formation was deposited.

* The pectens were kindly identified by Dr A. Beu, paleontologist with the New Zealand Geological Survey, Lower Hutt.
3. There is no apparent angular unconformity between the Brunswick Formation and the underlying beds correlated with the Landguard Formation, although the two wave cut surfaces each record an erosional hiatus.

Comparison of my stratigraphic log of the Warrengate section with that of Fleming (1953) - see Fig. 4.5) - suggests that the shell-grit collection (GS4240) made by Fleming, and attributed to the Brunswick Formation by him, actually comes from an horizon below the base of the Brunswick. The horizon from which he made this collection is apparently no longer exposed. The "leached Upper Castlecliff Shellbed" which Fleming named as directly underlying the Brunswick Formation sensu lato is here correlated with the Landguard Formation.

Representative stratigraphic sections through the Brunswick Terrace, and their locations, are shown in Fig. 4.3, together with all surveyed wave cut surface heights.

4.5.6.5 The Terangian Stage:

The type section of the Terangian Stage was designated at Mt Jowett (N138/593917) near Wanganui, by Fleming (1953, p.105), where the Brunswick Pebbly Sand yielded a rich fossil fauna (faunal list in Fleming 1953, p.260-261). However, the section is now completely obscured, and it is here suggested that a quarry on Brunswick Road (N138/582940) some 3km northwest of Mt Jowett be adopted as the new type section of the Terangian Stage (see Figure 4.3, Plate 4.5).

4.5.7 NGARINO TERRACE AND FORMATION (Dickson et al 1974 and redefinition)
4.5.7.1 Type Section:

The type section was designated by Dickson et al (1974) as a large soil slip west of Kai-Iwi Valley Road at grid reference N137/442933 (see Fig. 4.6 and Map 1). However, since that time this section has become largely revegetated, and in 1976 it was only poorly exposed. Here propose a new type section which is a road cutting on Omahina Road at grid reference N130/217108. Exposure is excellent (Plate 4.7).

4.5.7.2 Name:

The Ngarino Terrace was named by Dickson et al (1974) after Ngarino Road which descends to its surface from the Brunswick Terrace near grid reference N137/445962.

4.5.7.3 Reference Sections:

N129/041153 Kakaramea Road Extension
N130/115138 Matuku Road
N130/214125 Omahina Road
N137/265067 Moumahaki Road
N137/313021 Karaka

4.5.7.4 Nature and distribution:

The Ngarino Terrace is bounded at its inner edge by a high fossil cliff which separates it from the Brunswick Terrace (Plate 2.9). On the other hand, the fossil cliff at its outer edge is much more subtle, and is often difficult to locate. Dickson et al (1974) described this lower Rapanui/Ngarino cliff near N137/442929 where the height difference in wave cut surfaces above and below the cliff was reported as 12m (cf. estimate of 10m made in this work). West of Waverley on Karahaki Road, the height difference in wave cut surfaces reaches 15m near N130/128100; however, at most other localities it
appears to be considerably less (cf. the impressive differences in elevation of wave cut surfaces at the Ngarino/Brunswick cliff, which reach a maximum of c.50m inland from Waverley near grid reference N130/230121).

The distribution of the Ngarino Terrace, partly inferred where indicated is shown in Fig. 4.6 and Map 1. Representative sections are also shown in Fig. 4.6, together with surveyed wave cut surface heights. Dickson et al (1974) reported an average thickness of c.10m near Kai-Iwi for the cover beds of the Ngarino Terrace. Measurements in this work, over a wider area, indicate that c.15m is probably a more representative figure for the terrace sequence as a whole.

The Ngarino Formation was proposed by Dickson et al (1974) for the sedimentary cover of the Ngarino Terrace. Marine pebbly sands at the base of the formation were named the Sherwood Sand Member. In this work it is here proposed to redefine the Ngarino Formation to include only the marine sediments overlying the Ngarino wave cut surface. The name Sherwood Sand Member will lapse.

At the new type section on Omahina Road (Fig. 4.6), the Ngarino Formation consists of 6m of laminated pebbly sands containing abundant mafic minerals. Andesite pebbles form a basal conglomerate resting on an erosional unconformity, and decrease in abundance upwards. These marine sediments are conformably overlain by laminated micaceous silts which are interpreted to be estuarine. Overlying the silts is 5m of weathered andesitic aheses. One of these is a particularly prominent red-brown ash with Fe/Mn nodules, which has also been recognised at two sections within the Rapanui Terrace (N137/150088 on Kohi Road and N129/059120 on Otauto Road). This distinctive layer was
named the Parao Loess by Wilde (1979), who correlated it to several other sections; however, its degree of weathering makes genetic interpretation difficult. On the basis of its dark colour (which is similar to many weathered andesitic tephras further west, but different from pale grey-brown loesses east of Wanganui), I have preferred to refer to it as a weathered ash.

No dunesand, except for a thin layer some 10cm thick near the top of the exposure (Fig. 4.6) is present at the Omahina Road type section. However, thick, steeply cross-bedded, mafic sands are present at other localities (Fig. 4.6). On a farm track near Omahina Road at grid reference N130/214125, a thin, distinctive, pale coloured (rhyolitic?) ash rich in ferromagnesian minerals is present beneath dunesands and directly overlying marine sands. This ash may correlate with weathered ashy silts immediately overlying marine sands on Kakaramea Road Extension (grid reference N129/041153) - see Fig. 4.6).

No marine fossils have been observed, in this study, within the Ngarino Formation at any locality, although the wave cut surface is locally bored. Dickson et al (1974) reported friable, shallow water fossils including mollusc valves at one locality.

4.5.8 RAPANUI TERRACE AND FORMATION (Fleming 1953, and redefinition)

4.5.8.1 Type Section

The type section was designated by Fleming (1953 p.262) as: "the sediments forming the upper part of the section exposed in the sea cliff for 170 chains south-east of the mouth of Omapu Creek on the coast north-west of Castlecliff" ie. between grid references N137/802911 and N137/481891.
4.5.8.2 Name:

After the settlement of Rapanui 1.5km inland from the mouth of Omapu Stream.

4.5.8.3 Reference Sections:

N129/793269 Ohawe Beach
N129/793269 Waihi Beach
N129/944208 Manawapou River
N129/058119 Otauto Road
N137/126096 Karahaki Road
N137/150088 Kohi Road
N137/350952 Ototoka West Beach
N137/429926 Okehu Beach

4.5.8.4 Nature and distribution:

As indicated previously (p.99) the Rapanui Terrace is bounded at its inner edge by an inconspicuous cliff which separates it from the higher Ngarino Terrace. At its outer edge, the Rapanui Terrace is bounded by a low, obscured fossil cliff which separates it from the lower Inaha Terrace. On the coast some 1.5km south-east of the Manawapou River (N129/927185), the Rapanui/Inaha cliff is exposed in the modern cliff (Plate 4.9), and extends inland from there, at an angle of c.15° to the present shoreline (Map 1, Fig. 4.7). The Rapanui wave cut surface is exposed at an altitude of c.42m and the Inaha wave cut surface at the foot of the fossil cliff lies at an altitude of c.24m.

The distribution of the Rapanui Terrace is shown in Fig. 4.7 and Map 1. Representative stratigraphic sections are also shown in Fig. 4.7, together with surveyed wave cut surface heights.
Fleming (1953, p.263) stated that "the Rapanui Formation consists of the covering beds of the Rapanui Terrace". The Rapanui Formation is here redefined to include only the marine sediments overlying the Rapanui wave cut surface. The formation as redefined here is equivalent to the Rapanui Marine Sand Member of Fleming (1953). Fleming named the following members of the Rapanui Formation:

1. RAPANUI MARINE SAND, which was the basal, marine member of the formation. According to Fleming its lower boundary was an erosional unconformity which was usually overlain by "a think basal conglomerate, locally fossiliferous, overlain by 15ft of pebbly sand, similar to the Brunswick Pebby Sand but richer in magnetite, grading up locally to 10ft of well sorted, black beach sand" (Fleming 1953, p.263). Between the Tangahoe and Waingongoro River mouths, this unit (now Rapanui Formation sensu stricto) is continuously exposed and reaches up to 12m thickness. (In the same area, the Rapanui cover beds are in excess of 50m thickness owing largely to increased input of volcanic debris from the Taranaki volcanoes, and in particular laharc deposits, single units of which may be greater than 10m thick - see Plate 4.10)

A prominent shellbed, which is present almost continuously in coastal sections west of the Tangahoe River mouth, has been named the Denby Shellbed by Grace (1976). She also named the basal conglomerate, the Beachcroft Gravel as a new member of the Rapanui Formation sensu lato. Both these members now become named members of the formation sensu stricto.
Grant-Taylor & Beu (1974) have described the paleo-ecology of the Denby Shellbed at Waihi Stream (N129/820252) near Hawera. They concluded that a large estuary existed near the mouth of the present Waingongoro River at the time the shellbed was deposited, since 90% of mollusc species collected by them could tolerate less than normal ocean salinity. Grace (1976) has also studied the Denby Shellbed and concluded, on the basis of sedimentological and faunal evidence, that the Rapanui Marine Sand was deposited sub-tidally, i.e. of marine rather than beach origin. Furthermore, Grace (1974, 1976) concluded that sea temperatures could have been several degrees cooler than present at the time of deposition of the Denby Shellbed, and correlated it with an interstadial high sea level during the Otiran (last) Glaciation. These findings were contrary to those of Fleming (1953) who postulated sea temperatures several degrees warmer than present on the basis of faunal content of the Rapanui Marine Sand near Wairoa Stream (N137/170992). However, while this latter site has been traditionally assigned to the Rapannui Terrace, (e.g. Fleming 1953; Grace 1974, 1976) it is concluded in this work that it belongs to a younger terrace. This site is also the type section of the Oturian Stage (see discussion in 4.5.8.5 Oturian Stage, and also 4.5.10 Younger terraces and deposits.

2. WAIPUNA DELTA CONGLOMERATE, which according to Fleming (1953, p.263) "replaces the Rapanui Marine Sand near and south of the Wanganui [River]. It consists of gravel and pebbly sand deposited by the Wanganui and contains abundant hypersthene andesite among its constituent pebbles". This unit has not been studied in this work, and will not be considered further.
3. **RAPANUI LIGNITE**, which "overlies the Rapanui Marine Sand, locally with unconformity, and consists of andesitic ash, lagoonal clays, peaty lignite, fossil soils and fossil trees" (Fleming 1953, p.263). This unit appears to be widespread, and persistent, and is a useful distinguishing feature of the Rapanui cover beds at many localities. As defined here, the Rapanui Lignite is no longer a member of the Rapanui Formation, but overlies it at all localities where the lignite is present.

4. **RAPANUI DUNESAND**, "up to 25ft thick, overlies the Rapanui Lignite west of Wanganui River, and consists of dune-sand, locally displaying aeolian cross-bedding, similar in constitution to recent dunesand, but to some extent cemented by limonite and interstitial clay" (Fleming 1953, p.263) According to Wilde (1979) some of the deposits mapped by Fleming (1953) as Rapanui Dunesand can be demonstrated to be significantly younger than the sand exposed in a railway cutting at grid reference N137/440150, which Wilde (1979) designated as type section of the unit. He has named these younger sands separately (eg. Huxley Sand, Weraroa Sand, and Rangitatau Sand), but they will not be considered here.

5. **KAIWHARA ALLUVIUM**, which sometimes replaces the Rapanui Dunesand. In the Wanganui Valley it consists of "a thick sequence of pale silt and sand exposed at Landguard Bluff and for some miles to the north-east up Wanganui River and to the south-east along the coastal lowland" (Fleming 1953, p.263). It is apparently of fluvial origin.
At the eastern end of the type section, Fleming (1953, p.267) reported the following section (Fig. 4.8).

Recent dune sand.
(unconformity)
Aeolian iron sand and silty sand (Rapanui Dunesand) 10ft.
Lignite, from 16 to 30in thick with fossil wood near the base; locally absent (Rapanui Lignite).
Pale leached ashy clay (Rapanui Lignite) 2-3ft.
Bedded ironsand and pebbly sand (Rapanui Marine Sand), 30ft.
Conglomerate (Rapanui Marine Sand) 6 in.
(unconformity)
Shakespeare Cliff Sand (Castlecliffian Stage).

Fleming (1953) described palynological material from the Rapanui Lignite at this site which indicated climatic conditions little different from present.

Sections in coastal cliffs west of the Nukumarau Fault Zone, and particularly between the Patea and Waitotara Rivers) have traditionally been assigned to the Rapanui Terrace (Fleming 1953; Grace 1974, 1976). In this study they are assigned to a younger terrace.

4.5.8.5 Oturian Stage:

The type section of the Oturian Stage was designated near Wairoa Stream (N137/170992) by Fleming (1953, p.105), where the marine sand yielded a rich fossil fauna (see faunal list in Fleming (1953, p.272-274). The Oturian Stage was considered to represent the last interglacial by Fleming (1953) and Suggate (1965). On the basis of results presented here (see 4.5.10 Younger terraces and deposits), the Oturian Stage type section does not occur on the Rapanui Terrace.

4.5.9 INAHA TERRACE AND FORMATION (new name)

4.5.9.1 Type Section:

The type section is exposed in the coastal cliff some 200m west of the mouth of Inaha Stream at grid reference N129/748278 (Plate 4.11, Fig. 4.9).

4.5.9.2 Name:

After Inaha Stream which is entrenched below the terrace surface, and enters the sea 200m east of the type section.
4.5.9.3 Reference Sections:
N136/045032 Patea
N136/092064 Elslea Road
N137/104063 Rangikura

4.5.9.4 Nature and distribution:
The distribution of the Inaha Terrace is shown in Fig. 4.8, together with representative stratigraphic sections.

The Inaha Terrace was first recognised west of the Waingongoro River mouth by Aitken (in press), and subsequently by McGlone et al (1979). A marked, linear topographic rise extends inland from near the mouth of the Waingongoro River (Plate 4.12) and may be traced northwest as far as Inaha Settlement - a total distance of c.3km. This topographic rise, which separates two subhorizontal landsurfaces, and which is reminiscent of the surface expression of the Ngarino cliff, is thought to represent the surface expression of the buried Inaha cliff. Lack of suitable exposures unfortunately precludes mapping and surveying of height differences in wave cut surfaces along this cliff line.

East of the Waingongoro River, the Inaha Terrace has apparently been completely removed by subsequent marine erosion, and the older Rapanui Terrace is exposed in coastal cliffs as far east as the mouth of the Manawapou River (Fig. 4.8, Map 1). Immediately east of the Waingongoro River mouth, the Rapanui wave cut surface is 12m above HWM and increases in altitude to the east (Fig. 4.8, Map 1). Immediately west of the Waingongoro, the Inaha wave cut surface is 8m above HWM and decreases in altitude in coastal exposures to the west. At grid reference N129/745278 the Inaha wave cut surface dips below present beach level. This dip in the Inaha wave cut surface is
unlikely to be tectonic, but rather, is related to increasing distance from the Inaha strandline, which is oblique to the present coastline. The Inaha cliff is inferred to have intersected the present coastline at the mouth of the Waingongoro.

The Inaha Formation is here named to include all marine sediments overlying the Inaha wave cut surface. The Inaha Terrace is named as the terrace which truncates the outer edge of the Rapanui Terrace, and includes all deposits and features above and including the Inaha wave cut surface. From a point some 1.5km south of the Manawapou River mouth (N129/927185) where the Rapanui/Inaha cliff is exposed in the modern cliff (Plate 4.8), the Inaha Terrace is exposed continuously in coastal cliffs as far east as the Patea River mouth, and for several hundred metres beyond (Map 1, Fig. 4.8). Between the mouth of the Whenuakura River (N136/075046) and Wainui Beach at grid reference N137/290942, the Inaha Terrace is not exposed at the coast. The deposits here are assigned to a younger unnamed terrace, which is inferred to be separated from the Inaha Terrace by a low, obscured, fossil cliff some 1.5km inland of the present coastline (Fig. 4.10). The Inaha Terrace is inferred to be exposed in coastal cliffs between Wainui Beach and the Nukumaru Fault Zone (Fig. 4.8); however, poor exposures prevent any meaningful examination of this c.3km long portion of the coast. The Inaha Terrace is not present east of the Nukumaru Fault Zone.

At its type section (N129/748278) the Inaha Terrace consists of a c.36m thick (Plate 4.11, Fig. 4.9) sequence of predominantly interbedded laharic debris and carbonaceous deposits overlying 2m of laminated, coarse, andesitic marine sands. The basal portion of these marine sands contains andesitic pebbles and boulders, many of which have been bored by rock boring molluscs, one sample of which has been
identified as Lithophaga (Zelithophaga) truncata (Gray); (V.E. Neall pers comm (1979)). No other marine fossils are evident, except for poorly preserved bivalve casts in silty sand approximately 1km east of the type section. A 1m thick, prominent lignite between 4 and 5m above the wave cut surface has been informally referred to as the Manaia Lignite by McGlone et al (1979). Higher up in the section, a 10m thick lahar unit records a large collapse of probably the ancestral Mt Egmont (McGlone et al 1979).

4.5.10 YOUNGER MARINE TERRACES AND DEPOSITS

Between the mouth of the Whenuakura River (N136/075046) and Wainui Beach (N137/290942) an unnamed terrace younger than the Inaha is inferred to be present. Coastal exposures are more or less continuous, but inland there are no useful exposures. Wave cut surface elevations can therefore only be determined along the present coastline (Fig. 4.10). Representative sections are shown in Fig. 4.10. Cover bed thicknesses are variable up to 12m.

Between Snapper Point (N137/153993) and the mouth of Wairoa Stream (N137/167992) Fleming (1953, p.264) described the following generalised section which he attributed to the Rapanui Formation (sensu lato):

- 6ft Consolidated Rapanui Dunesand.
- 9in Fossil soil and lignite.
- 2ft Pale lacustrine clay.
- 8ft Sand with muddy partings.
- 6in Current bedded sand and conglomerate.
- 8ft Sand containing Arachnoides and rare Barbatia valves fossiliferous marine conglomerate.
   (unconformity)

Whenuakura Group.

Some 300m east of the mouth of Wairoa Stream at grid reference N137/170992 Fleming (1953) reported a rich fauna from marine sediments overlying the wave cut surface. This latter locality was designated by Fleming as the type locality of the Oturian Stage, which he correlated with the last interglacial. From this work it is concluded that this type section does not belong to the Rapanui Formation, but rather
to terrace deposits within the younger unnamed terrace which post-dates the Inaha.

Fleming (1953, p.264) noted that "east of Wairoa Stream mouth, cliff top sections show unconformity between the Kaiwhara Alluvium...and the underlying Rapanui Marine Sand". In this study three $^14C$ dates were obtained on wood samples from the beds referred to by Fleming as Kaiwhara Alluvium (see Plate 4.13, Fig. 4.10, Waverley Beach Section) and indicate that much of the deposit is Holocene in age. At grid reference N137/185985, sample BJP-040 from a peat lens some 4-5m above HWL have an age of 6200±100 years BP (ANU-1891). Sample BJP-039 which gave an age of 7300±110 years BP (ANU-1890) was collected from a carbonaceous silt deposit extending below the present HWL. This deposit, which contains tree stumps in growth position lies within a small fossil stream valley which has been infilled with peat and dunesands (Fig. 4.10). A large stump on the modern beach face at the same locality is interpreted to represent the seaward extension of the deposit. At the time this tree grew relative sea level must have been at least 2m lower than present ie. there has been a relative rise in sea level of at least 2m since c.7300 years BP. Sample BJP-037, from a small channel infill of grey silts at present HWL some 500m to the east (N137/192983) gave an age of 7500±110 years BP (ANU-1888).

Some 5km to the east of the Waverley Beach site, Fleming (1953, 1957b) reported submerged tree stumps in growth position at the mouth of the Waitotara River (N137/237963). A sample from one of these yielded a $^14C$ age of 1020±60 years BP (Fleming 1957b), and indicates a relative rise in sea level of at least 2m since that time, despite a long history of uplift along this coast (cf. the terrace sequence itself).
Approximately 500m west of the mouth of the Whenuakura River at grid reference N136/064047, an erosional unconformity is exposed at c.2m above present HWM. It is overlain by laminated andesitic sand containing a basal conglomerate. This exposure is interpreted as a small marine terrace remnant which is younger than the post-Inaha Terrace to the east (the wave cut surface on the eastern side of the Whenuakura at grid reference (N136/077040 is c.6m above HWM). Age is uncertain, but is interpreted to be pre-Holocene (see discussion p.131).

Other low terraces are present near major river mouths between Hawera and Wanganui, but most are clearly of fluvial origin. Several samples have been C\(^{14}\) dated from terraces near the mouth of the Manawapou River (Fig. 4.10), and indicate a complex history of fluvial development both during and prior to the Holocene. An undulating erosional surface between 3 and 8m above HWM occurs within a low terrace between the mouths of the Tangahoe and Manawapou Rivers (Fig. 4.10). Laminated and trough bedded pebbly sands, which are andesitic in part, overlie this surface. This terrace is possibly, in part at least, of marine origin, and its altitude suggests possible correlation with the post-Inaha Terrace.

A small terrace remnant occurs near the mouth of Mowhanau Stream (N137/440917) and was described by Fleming (1953, p.23, Fig.3) as being of fluvial origin. Laminated pebbly sands (with basal conglomerate) overlie a planar, sub-horizontal erosional surface some 5m above HWM and are suggestive of a marine origin. Owing to the steepness of the exposure it was not possible to examine the terrace more closely; correlation is uncertain.
4.6 CONCLUDING REMARKS

In this chapter, eight marine terraces have been described and named between Hawera and Wanganui as follows, from youngest to oldest: Inaha, Rapanui, Ngarino, Brunswick, Upper Brunswick, Kaiatea III, Kaiatea II and Kaiatea I. Of these, one (the Upper Brunswick Terrace) has not been previously recognised, whilst four (Inaha, Kaiatea I, Kaiatea II, Kaiatea III) have not been previously adequately defined. In addition, at least one and possibly two marine terraces younger than the Inaha Terrace are inferred to be present. The Oturian Stage type section, previously thought to occur within the Rapanui Terrace, is correlated with the older of these two post-Inaha terraces. At least two marine terraces older than the Kaiatea I Terrace are inferred to be present, but have not been named at this stage.

An important stratigraphic marker, the Omahina Tephra (fission-track dated at c.370Ka BP) has been demonstrated to overlie wave cut surfaces within terraces older than, and including, the Kaiatea III Terrace. The cutting of these wave cut surfaces must therefore predate the fall of the Omahina Tephra. The absence of the Omahina Tephra from younger terraces implies that all wave cut surfaces younger than that of the Kaiatea III Terrace must post-date its fall. If correlation of the Omahina Tephra with the Rangitawa Pumice (which occurs within Upper Castlecliffian marine sediments at its type section) is accepted, a Castlecliffian age for the Kaiatea III and older terraces is indicated. Even if the correlation is rejected, the fission track age of the Omahina Tephra (c370Ka BP) still supports a Castlecliffian age for these terraces (Boellstorff & Te Punga (1977) estimated an age of c.340Ka BP for the top of the Castlecliffian Stage).
The Brunswick Formation (sensu stricto) has been demonstrated to stratigraphically overlie the Landguard Formation, and is therefore referred to the Hawera Series. The Hawera Series/Castlecliffian Stage boundary is thus inferred to fall between the cutting of the Brunswick and Kaiatea III wave cut surfaces. If the second wave cut surface at the Warrengate Section (N138/732815) is correlated with the Upper Brunswick wave cut surface, then the boundary is inferred to be between the cutting of the Upper Brunswick and Kaiatea III wave cut surfaces.

As indicated previously (p.71) all terraces of the South Taranaki-Wanganui sequence have been traditionally assigned to the Hawera Series. Part of the reasoning for this was that the Hawera Series/Castlecliffian Stage boundary was seen to represent a regional unconformity (see Fleming 1953, p.247 for example), separating deposits clearly related to the present landscape (Upper Quaternary) from underlying deposits which were in most instances quite discordant with present topography and more closely similar to those of the Pliocene (Lower Quaternary). However, stratigraphy and dating in this study do not support this interpretation. In particular shallow marine and littoral sediments within the Kaiatea III and older terraces are inferred to represent the shoreline equivalents of Castlecliffian marine sediments within the South Wanganui Basin, and clearly indicates a more complex history than this "regional unconformity" model implies. The deformation history of the terrace sequence is considered in the following chapter.
CHAPTER 5

INTERPRETATION AND DATING OF THE TERRACE SEQUENCE

In this chapter, tectonic, eustatic and climatic interpretation of the terrace stratigraphy is attempted. Age estimates of the terraces are presented, and correlation within New Zealand is discussed. Correlation is also attempted with some of the more important long Quaternary records from other parts of the world.

5.1 TERRACE DEFORMATION : PRINCIPLES

As indicated previously (p.70) the terrace sequence owes its origin to successive eustatic oscillations superimposed on a vertically rising landmass. This uplift apparently is linked with the emergence of the South Wanganui Basin in the Upper Quaternary.

To analyse the pattern of uplift, two elementary types of deformation are first considered: doming uplift and hinging uplift (Fig. 5.1). In the case of doming uplift, rates of vertical rise increase radially inwards towards a central region of maximum uplift, while for hinging uplift, rates of vertical rise increase linearly away from a line of no vertical uplift (zero uplift isobase). Doming-type uplift resembles the isostatic uplift pattern of the Fennoscandia and Laurentide areas which has occurred following deglaciation during the past c.20Ka. Isostatic deformation of Lake Bonneville strandlines since the lake drained some 12Ka BP (Bloom 1967) is also similar. Hinging uplift, or simple tilt, is approximated on the flanks of active folds and also resembles the pattern of deformation associated with changing water loads on linear continental shelves (hydroisostatic deformation).
With reference to a modern coastline, deformation patterns associated with each of the above uplift models are idealised in Fig. 5.1. Rates of uplift, relative to two orthogonal traverses (shore-normal and shore-parallel) are shown for each, and illustrate the essential differences in deformation pattern. However, despite these differences, it is clear from Fig. 5.1 that at least over short distances, in the case of doming uplift, the actual shore-normal uplift pattern can be reasonably approximated by a hinging-type model. This observation has important ramifications in the context of the South Taranaki marine terrace sequence, and will be discussed in the following sections. For simplicity, "shore-normal" will be taken to mean "in a direction approximately normal to the trend of the terrace fronts" (ie. ascending or descending the terraces), and "shore-parallel" will mean "in a direction along any single terrace".

The actual pattern of shore-parallel deformation in South Taranaki is shown in Fig. 5.2, in which wave cut surface elevations and inferred strandlines have been plotted for each terrace (projected onto a line joinging Hawera and Wanganui railway stations. The deformation pattern is similar for all strandlines and depicts a gentle, doming structure with maximum uplift rates near Waverley (Fig. 5.2). Wave cut surfaces and strandlines are vertically offset along the Moumahaki, Ridge Rd and Waitotara Faults, as well as along the Nukumaru Fault Zone (Fig. 5.2).

The Nukumaru Fault Zone, Waitotara Fault and Moumahaki Fault were previously described and mapped by Fleming (1953), while the Ridge Rd Fault is identified for the first time in this work. Surface traces are shown on Map 1. All appear to be normal faults; no component of horizontal movement has been identified. Strikes are roughly NNE-SSW. Fault offsets relative to individual terrace wave
cut surfaces are listed in Table 5.1. The Moumahaki Fault has been observed to offset the Upper Brunswick, Brunswick and Ngarino wave cut surfaces each by c.40m (to nearest 5m). No offset on the Rapanui Terrace is proven: exposure is poor, but surface morphology suggests that the Rapanui Terrace may truncate the Moumahaki Fault. The Ridge Rd Fault (named in this work after Ridge Road which it crosses over near N130/307124), offsets the Brunswick wave cut surface by c.30m, and the Kaiatea III wave cut surface by c.80m. The Waitotara Fault offsets the Brunswick and Ngarino wave cut surfaces by c.15m, but has not been demonstrated to offset other wave cut surfaces. The Nukumaru Fault Zone was named by Fleming (1953) for a c.600m wide zone of faulting which runs inland from the coast (at N137/326954) east of the Waitotara River. It can be demonstrated to offset all terraces older than and including the Rapanui; vertical offset increases up the terrace flight (Table 5.1), suggesting continuous vertical movement throughout the period of terrace development.

As indicated above, although shore-parallel doming-type uplift is evidenced by the strandline deformation patterns (Fig. 5.2), locally this uplift may be approximated by hinging uplift along carefully selected transects. For example, between traverses 2 and 6 (Fig. 5.2), deformation of individual strandlines is approximately linear. This deformation (expressed as tilt in degrees of arc) can be calculated for each strandline using the formula

$$\tan^{-1}\left\{ \frac{(h_6-h_2)}{d} \right\}$$

where

- $H_6 =$ strandline height at traverse 6
- $H_2 =$ strandline height at traverse 2
- $d =$ horizontal distance between traverse 6 and traverse 2
- $= 20Km$
Tilts are listed in Table 5.1 (column 5), and indicate progressive tilting throughout the history of the terrace sequence. Assuming that this tilt has been occurring at a constant rate, the data indicate that the Ngarino strandline is about \( \frac{1}{2} \) as old as the Kaiatea III strandline, while the Kaiatea I strandline is about 1.5 times as old as the Kaiatea III. Thus, knowing the age of the Kaiatea III strandline for example, it would be possible to estimate ages for the other terrace strandlines (see 5.2 Geochronology).

Turning now to shore-normal deformation, a simple tally of mean wave cut surface slopes derived from altimeter survey data (Table 5.1 Column 6) once again indicates increasing tilt up to terrace flight and progressive deformation throughout the history of terrace development. As with shore-parallel deformation, by assuming a constant rate of tilting and knowing the age of one terrace, it would be possible to estimate ages for other terraces. However, since the wave cut surfaces were not originally horizontal (in a shore-normal direction), correction would have to be made for initial slope.

Further analysis of shore-normal deformation is possible using height relationships between terrace strandlines on each of the 14 shore-normal transects shown in Fig. 5.2. These equally spaced shore-normal transects do not represent field traverses; however, data points are assumed to be evenly distributed for the sake of simplicity, so that strandline heights can be assumed to be equally accurate on each. Strandline heights are given in Table 5.2 for each transect. Strandline height ratios have been calculated for each terrace relative to the Kaiatea III strandline, for each of the 14 shore-normal transects (Table 5.3). Although strandline height ratios remain essentially constant along the terrace sequence, such constancy cannot be interpreted to indicate uniform uplift rates along each transect.
Consider first the simplest case: a terrace sequence with many shore-normal traverses, and uplift rates constant along each traverse, but varying from traverse to traverse. Assuming past sea levels associated with each terrace were similar in absolute (geoidal) terms, this situation is depicted in Fig. 5.3A and has been called a "relation diagram" by Wellman (1971). In this case strandline height ratios are constant for any two given strandlines and are equal to the ratio of their ages. This model is clearly inappropriate for the South Taranaki terraces in view of the demonstrated increase in wave cut surface tilts up to terrace flight (see Table 5.1 column 6) i.e. shore-normal tilting.

The next simplest case is to assume a hinging-type uplift, with constant tilt along each traverse, but with mean uplift rates varying between traverses (Fig. 5.3B). Once again strandline height ratios for any two given strandlines are constant, even if the hinge line is not shore-parallel (Fig. 5.3D); however, in this case height ratios do not represent age ratios. Note that when traverses are along the strike of the hinging surface, the uniform uplift model (case 1) applies. While this hinging-type uplift model may be locally applicable in South Taranaki (see also shore-parallel tilting between traverse 2 and 6 in Fig. 5.2), it cannot continue indefinitely cf. Mt Ruapehu in the centre of the North Island.

Thus a more likely model of shore-normal deformation in South Taranaki is more complicated again, and may approximate a doming-type uplift in which the rate of increase in uplift rates decreases inland along each traverse. In this case height ratios are only constant if the doming pattern is similar for each traverse (i.e. parallel traverse lines on relation diagram) - see Fig. 5.3C. Once again, height ratios do not represent age ratios. (Mean strandline height ratios are given in Table 5.2 column 7).
To summarize, doming-type uplift can be demonstrated to be occurring on the terrace sequence. Over short distances, however, this may be approximated by hinging uplift. In the following section ages of the terraces are estimated, which allows further discussion of the timing and rates of terrace deformation.

5.2 GEOCHRONOLOGY

5.2.1 STRATIGRAPHIC SIGNIFICANCE OF THE OMAHINA TEPHRA

As indicated previously (p.80) the Omahina Tephra closely overlies marine sediments within the Kaiatea III Terrace at many localities (Fig. 4.1, Appendix 2). The deposition of the Kaiatea III Formation and cutting of the Kaiatea III wave cut surface have therefore been inferred to closely precede its fall (p.80). Unfortunately it is difficult to estimate how much time elapsed between emergence of the Kaiatea III Formation and the fall of the Omahina Tephra. One crude method, however, is to assume that, following emergence, non-marine cover bed thickness increased linearly with time at any one locality. Thus by measuring the present depth of burial of the tephra at a site, and comparing it with the total non-marine cover bed thickness at the same site, and accepting an age of 370Ka BP for the fall of the Omahina Tephra, an estimate can be made for the time of emergence. Using data from the Ridge Rd and Omahina Rd sections (Fig. 4.1) in this way, the time of emergence can be estimated at each site as 390 and 400Ka BP respectively.

At Meremere, east of Hawera (Locality 6 Fig. 4.1), a c.4m thick tephric lignite (informally referred to as the Meremere Lignite - see p.86) occurs beneath the Omahina Tephra in cover beds of the Kaiatea II Terrace. Pollen samples from this site indicate a gradual
transition from podocarp/rata forest to beech forest, which may be interpreted as the dying phase of an interglacial period (M.S. McGlone pers comm 1980). Lying as it does between the Omahina Tephra and the Kaiatea II wave cut surface (Fig. 4.1), this lignite probably relates to the time of emergence of the Kaiatea III Terrace. Simple extrapolation of cover bed accumulation rates (as above) at this site indicates an age of c.400Ka BP for the top of the lignite, which is in good agreement with the estimated times of emergence for the Kaiatea III Formation at both Ridge Rd and Omahina Rd sections.

Hence, emergence of the Kaiatea III Terrace is assumed to have been occurring at c.400Ka BP. The culmination of the marine transgression which cut the Kaiatea III wave cut surface must therefore have occurred immediately prior to this. In the absence of any firm evidence as to the exact timing of this culmination, an age of c.400Ka BP is also adopted. Even if the Kaiatea III strandline was last active up to 20Ka prior to emergence of the Omahina Rd and Ridge Rd sites, an error of not more than 5% in the age estimate is introduced*. A mean age of c.400Ka BP is also adopted for the Meremere Lignite.

5.2.2 AMINO ACID RACEMISATION DATES

The amino acid racemisation ratios (alloisoleucine/isoleucine) are calibrated to the Meremere Lignite, for which a 400Ka age is adopted from its relationship to the Omahina Tephra as described above. Four wood samples from this lignite (BJP-026, 027, 028 and 029) were used to calibrate the dating method (see Chapter 3 for discussion, Table 3.4 for age summary, and Fig. 3.11 for dating

---

* Wave cut surface dips, initially less than 0.4° (see Table 5.1 column 6), distance of the Ridge Rd and Omahina Rd sites seaward of the strandline (c.2Km), and estimated rates of eustatic change during major transgression/regression cycles in the Quaternary (cf. sea level curve of Chappell 1974b), suggest a figure of less than 10Ka for the culmination/emergence lag time.
curve). This has enabled samples to be dated from the Brunswick, Ngarino, Rapanui, Inaha and post-Inaha Terraces; no samples have been dated from the Upper Brunswick and older terraces.

Five wood samples from the non-marine cover beds of the Brunswick Terrace BJP-020X, 021B, 022X, 044/44B and 045 (see Figs. 3.11, 4.4 and Table 3.4) yield ages that range from 120Ka BP up to 195Ka BP. Since all these samples stratigraphically overlie the Brunswick Formation, the Brunswick strandline may be assigned an age of greater than c.200Ka BP. In addition, Boellstorff & Te Punga (1977) have estimated an age of 340Ka BP for the top of the Castlecliffian Stage; since the Brunswick Formation stratigraphically overlies the Landguard Formation at the Warrengate section (Figs. 4.4 and 4.6), an upper age limit of c.340Ka BP is assigned to the Brunswick strandline. Thus, the Brunswick strandline is inferred to have been active sometime between 200 and 340Ka BP; more precise estimation is not possible at this stage.

Two samples, BJP-049 and 05Ü (Figs. 3.11, 4.7 and Table 3.4) from estuarine sediments conformably overlying the Ngarino Formation at separate localities give ages of 190 and 180Ka BP respectively. An age of about 200Ka BP is therefore adopted for the Ngarino strandline. Note here that on the basis of shore parallel tilting, the Ngarino strandline was previously (p.117) estimated to be roughly half the age of the Kaiatea III strandline. The agreement of these two age estimate relationships from separate sources is encouraging.

The most extensively dated terrace is the Rapanui Terrace. Eight wood samples have been dated from the non-marine cover beds of the Rapanui Terrace (see Table 5.4 for summary as well as Table 3.4
and Figs. 3.11 and 4.8). They range in age from 78 Ka BP (BJP-033) to 132 Ka BP (BJP-030F), and have a mean age of c.105 Ka BP. Omitting sample BJP-033 from the sample average gives a mean age of c.110 Ka BP for the remaining samples. Since all these samples (seven) lie stratigraphically low in the non-marine cover beds of the Rapanui Terrace, an age of somewhat greater than c.110 Ka BP is suggested for the Rapanui strandline. In addition, a sample (BJP-001) from a terrace of the Waingongoro River, incised below the Rapanui wave cut surface gives an age of c.74 Ka BP (cf. $^4$C age >35,200 (ANU-1880) – see Tables 3.3 and 5.4). A well dated, major, high sea level event between 120 and 135 Ka BP is well documented from U/Th dating of ancient coral reefs in many parts of the world (see Chappell & Veeh 1978a for discussion). This event, also identified in oxygen isotope studies of deep sea cores, is referred to as Termination II (Broeker & Van Donk 1970; Shackleton & Opdyke 1973), and is generally correlated with the last interglacial sensu stricto (see Suggate 1974). Culmination of this major transgression is thought to have occurred at c.120 Ka BP when sea level (low water level datum) lay between 5 and 8 m above present sea level (Chappell & Veeh 1978a). Correlation of the Rapanui strandline with this event is made here, and the Rapanui strandline is inferred to have been last active c.120 Ka BP.

Four samples have been dated from the Inaha Terrace west of the Waingongoro River. Two of these (BJP-014X and 015X), from the upper portion of the type section (Fig. 4.10) give ages of 31 and 23 Ka BP respectively, and are in fair agreement with a $^4$C determination of 33,000±1100 years BP (ANU-1887) from the same horizon. Sample BJP-010X from the Manaia Lignite near the base of the type section (Fig. 4.10) gave an age of 93 Ka BP, while BJP-060 from the Inaha Formation (marine) some 1.5 km to the east gave an age of 73 Ka BP. In both these latter samples there was considerable interference from
overlapping peaks in the amino acid auto-analyser chromatograms; consequently the age estimates are of somewhat lower reliability. An age of c.80-100Ka BP seems likely, however, for the Inaha strandline.

Three samples from the post-Inaha Terrace at Waverley Beach were analysed (see Fig. 4.11). Two of these (BJP-037 and 040) are of undoubted Holocene age ($^{14}C$ ages 7500±110 and 6200±100 years BP respectively - see Table 3.3), and yet give racemisation age estimates of 15 and 27Ka BP respectively. While these racemisation age estimates are in seemingly poor agreement with the $^{14}C$ ages, it is worth pointing out that:

(a) The difficulty in measuring the small amount of D-isomer present in such young samples makes calculation of D/L ratios (and ages) subject to large errors.

(b) In situ temperatures experienced by such young samples are likely to be well above the long term (c.400Ka) average, and hence D/L ratios will be anomalously high - see discussion Chapter 3.

The third sample, BJP-041, from marine sediments overlying the post-Inaha wave cut surface at Waverley Beach (Fig. 4.11) gave an age of 62Ka BP. A similar age for the post-Inaha strandline is inferred.

5.2.3 EVALUATION OF DEFORMATION AND GEOCHRONOLOGIC DATA

As indicated previously (5.2 Terrace Deformation), terrace ages can be estimated using the deformation data in Table 5.1, assuming that:

(1) the pattern of terrace deformation has remained constant,

(2) the age of one terrace is independently known, and in some cases

(3) past high sea levels, during which the terraces were cut, all were similar to the present level.

Assuming an age of 400Ka BP for the Kaiatea III strandline, and subject to the other constraints above and listed in Table 5.4, age estimates are now made for the other terraces using
(a) shore-parallel tilt data (strandlines) - column 5 Table 5.1.
(b) shore-normal tilt data (wave cut surfaces) column 6 Table 5.1.
(c) vertical offset on the Nukumaru Fault Zone - column 4 Table 5.1.

These are listed in Table 5.5 and are compared with the independently derived age estimates based on fission track and amino acid racemisation data. Each of the deformation data sets yields terrace age estimates which are in fair agreement with each other, and also in moderate agreement with the fission track/racemisation age estimates. Such agreement is a little surprising in view of the fact that each of the three deformation models is at best only a crude estimate of the actual deformation.

Also listed in Table 5.5 for comparison are age estimates based on mean strandline height ratios (see column 7 Table 5.1) relative to the Kaiatea III strandline. However, it has been demonstrated previously that shore-normal tilting is present (cf. wave cut surface tilts), so that age estimates based on height ratios (which assumes a constant uplift rate for any shore-normal transect), will be in increasing error with increasing horizontal distance from the reference (Kaiatea III) strandline. In particular, age estimates will be increasingly too low for progressively younger terraces.

In the case of the Nukumaru Fault Zone vertical offset data (Table 5.1 column 4), derived age estimates based on a constant rate of fault movement (Table 5.5) are subject to large error bounds owing to the lack of precision in the primary data (±5m for each strandline). Despite this, fair agreement with the other model ages is apparent.

A vertical offset of 30m on the 400Ka Kaiatea III strandline indicates a mean rate of relative vertical movement of 0.075mm/year or 7.5cm/Ka for the past 400Ka. Fleming (1953, p.291) recorded a
maximum throw of about 500' (152m) on the Kuranui Limestone (of Hautanian age) on the Nukumaru Fault Zone. Recent fission track dates from the Ohingaiti Sand, which is at the base of the overlying Marahauan Substage, vary between 1.5my (glass: Seward 1974, 1976) and 1.78±0.44my (zircon: Seward 1979). Assuming an age of c.1.9my for the Kuranui Limestone, and a 152m offset, yields an average rate of vertical movement of c.8cm/Ka, which compares quite closely with the c.7.5cm/Ka rate derived from the Kaiatea III strandline. These very low, apparently constant, rates of vertical offset on the Nukumaru Fault Zone are interesting as it has been suggested by some workers (e.g. Suggate 1978b, p.686) that the Fault Zone is the southerly continuation of intense faulting in the Taupo Volcanic Zone, and links further south with the Alpine Fault System in the South Island. Furthermore, little or no horizontal displacement is evident. Thus, activity on the Nukumaru Fault Zone has remained anomalously low during the last c.2my considering its tectonic position, and no support is found for the notion that the western boundary of the Hawkes Bay Coastal Microplate passes along it (cf. Ballance 1976).

In the case of the shore-parallel deformation model, in which doming uplift has been approximated by hinging uplift, the mean rate of tilt, based on the Kaiatea III strandline, amounts to 0.15°/400Ka or c.7n.rad/year or c.1°/2.5my. Potentially this is thought to give the best estimates of strandline ages of the four models.

In the case of shore-normal deformation, approximated as a hinging uplift, the mean tilt rate using the Kaiatea III wave cut
surface (0.88° dip, 0.27° estimated initial dip*, 400Ka age assumed) amounts to 0.61°/400Ka or c.27n.rad/yr or 1°/660Ka ie. approximately 4 times as great as the shore-parallel rate. This shore-normal tilt rate may be compared with a mean tilt rate of c.60n.rad/year determined by Milne (1976) from his study of river terrace deformation in the Rangitikei Valley some 50Km to the east of Wanganui. (Terrace profiles up the Rangitikei Valley (Milne 1973b) indicate a convergence point for the terraces near present river level some 10-15Km upstream of the mouth, and inland of which the terraces become progressively higher). Unfortunately, although Fleming (1953) noted a similar tilting of river terrace profiles along the Wanganui, Whangaehu and Turakina Rivers, a lack of dates and accurate survey information preclude quantification. However, terrace profiles up these rivers (Fleming 1953, Fig. 20) indicate that their convergence point lies some distance seaward of their present mouths. ie. the zero uplift isobase is offshore for these rivers (cf. Rangitikei River Valley where it is onshore).

To summarise, each of the above crude deformation models yields terrace age estimates that are in moderate agreement with each other and with independent age estimates and constraints (fission track and racemisation dating methods). Deformation rates may be calculated accordingly. In the next sections, these age estimates will be further refined and separation of eustatic and tectonic factors attempted.

* Linear regression of maximum wave cut surface height(X) and wave cut surface dip(Y) for each terrace yields a relation of the form $Y = 0.27 + 0.0024X$, $R^2 = 0.99$. Assuming that maximum height is proportional to age, the intercept value (0.27) is close to the initial wave cut surface.
5.3 EUSTATIC CHANGES

5.3.1 INTRODUCTION

Uplifted marine terraces similar to the South Taranaki-Wanganui sequence have been described in California (Kern 1977, Karrow & Bada 1980, Woods 1980) and Japan (Machida 1975). Chronology in these places depends on fission track dating (Japan) and a combination of U/Th and amino acid racemisation dating (California). Better dated coral terraces, where chronology is based on extensive U/Th dating, occur in New Guinea (Bloom et al. 1974, Chappell 1974b), Atauro and Timor (Chappell & Veeh 1978a), Barbados (Broeker et al. 1968, James et al. 1971, Bender et al. 1979) and the Ryukyu Islands (Konishi et al. 1970); these provide a more complete chronology of late Quaternary sea level changes, in particular the timing of relative high sea level events. Table 5.6 summarises data from the above mentioned places; tentative correlations are also shown. Unfortunately, in nearly all cases, the separation of eustatic and tectonic factors is complicated by non-uniformity of tectonic processes and rates. Consequently there is some disagreement as to the absolute heights (relative to present) of the sea level events which produced the various terraces, particularly prior to c.120Ka BP.

In New Guinea (Huon Peninsula), which has the most complete record because of high uplift rates, the terrace flights back to 120Ka BP are steep with individual reef terraces being narrow; hence errors due to shore-normal tilt are argued to be small compared with uplift effects. However, terrace height ratios vary on transects normal to the coast (Chappell 1974b, p.566), indicating that uplift rates on single transects have not remained constant, although Bloom et al. (1974, p.202) concluded that "...the assumption of constant uplift is the best assumption now available". Again, at Atauro Island, which
is an extinct late Tertiary volcano north of Timor, uniformity of uplift seems likely, as uplift rates appear to be regional and extend across to Timor; hence shore-normal tilt error is assumed to be negligible (Chappell & Veeh 1978a).

The uplifted coral reef terraces of Barbados have been extensively dated using U/Th and He/U dating methods. Once again uplift rates have been assumed to be constant over short shore-normal transects (Broeker et al. 1968). Matthews (1973) demonstrated that uplift rates on two shore-normal transects had remained constant relative to each other, during the last c.120Ka. However, Bender et al. (1979) have argued that prior to c.120Ka BP uplift rates (in part of the island at least) were much greater than the rates after c.120Ka BP, while Mesolella et al. (1969, p.252, Fig. 3) have indicated considerable shore-parallel differential uplift. Consequently regional deformation patterns are likely to be considerably more complex than a simple uniform uplift model.

A synthesis sea level curve based on data from New Guinea and Atauro (after Chappell 1981), where separation of eustatic and tectonic factors appears to have been most successful, is given in Fig. 5.4. Dated high sea level events identified in Barbados (Bender et al. 1979) are also indicated for comparison. Also shown in Fig. 5.4 are $\delta^{18}O$ curves from (1) Broeker & Van Donk (1970 and (2) Shackleton & Opdyke (1973). Chronology in the deep sea cores relies heavily on constant sedimentation rates between points of known age eg. between the Brunhes/Matuyama boundary dated at c.730Ka BP (Mankinen & Dalrymple 1979) and present. Furthermore, the $\delta^{18}O$ record does not represent simply an ice volume (sea level) signal: sediment mixing can significantly smooth the record, while ocean temperature effects may also be present. Despite these limitations, glacial-interglacial
cycles in the deep sea core records show satisfactory agreement with the same broad $10^5$ year cycle evident in the New Guinea/Atauro sea level curve (Chappell 1981).

The sea level and $\delta^{18}O$ curves shown in Fig. 5.4, I consider to be the best available records of eustatic changes over the past c.700Ka. They are therefore used as a reference against which the eustatic record of the South Taranaki terraces may be compared.

5.3.2 **SEPARATION OF EUSTATIC AND TECTONIC FACTORS IN SOUTH TARANAKI**

The starting points for separation of tectonic and eustatic factors in South Taranaki are the Kaiatea III and Rapanui strandlines. The age of the Kaiatea III strandline is estimated in this work to be 400Ka BP. Sea level associated with this strandline is assumed to have been equal to present. On this basis, mean uplift rates have been calculated for the Kaiatea III strandline heights given in Table 5.2. These uplift rates, which vary between 0.48mm/year (Transect 1) and 0.66mm/year (Transects 6, 7, 8), are listed in Table 5.7. The Rapanui wave cut surface is assumed to have been formed by a high sea level event culminating at c.120Ka BP, with a paleosealevel of +5m relative to present (see p.122for discussion). On this basis mean uplift rates vary between 0.29mm/year (Transect 13) and 0.54mm/year (Transects 6, 7, 8) for the Rapanui strandline heights given in Table 5.2; calculated uplift rates are listed in Table 5.7. Note that in all cases, uplift rates for the Rapanui strandline are much less than those of the Kaiatea strandline on the same shore-normal traverse.

Based on a hinging uplift model, uplift rates would be expected to increase linearly from the Rapanui strandline inland to
the Kaiatea III strandline. On this basis model uplift rates are interpolated for the Ngarino, Brunswick and Upper Brunswick strandlines (Table 5.7) on each shore-normal traverse using the relation

\[ U_{ij} = U_{rj} + \frac{d_{ij}}{d_{kj}} (U_{kj} - U_{rj}) \]

- \( U_{ij} \) = uplift rate on intermediate strandline \( i \), traverse \( j \)
- \( U_{rj} \) = uplift rate on Rapanui strandline, traverse \( j \)
- \( U_{kj} \) = uplift rate on Kaiatea III strandline, traverse \( j \)
- \( d_{ij} \) = horizontal distance between Rapanui and intermediate strandlines, traverse \( j \).
- \( d_{kj} \) = horizontal distance between Rapanui and Kaiatea III strandlines, traverse \( j \).

Extrapolation of the hinging uplift model to terrace strandlines younger than the Rapanui and older than the Kaiatea III seems unwise. Firstly, as indicated previously (p. 118), hinging uplift cannot be extended inland indefinitely, and secondly, the hinging uplift model (as an approximation to doming-type uplift), is best restricted to short shore-normal traverses. On the other hand, the lack of suitable datums, i.e. strandlines of known age and absolute sea level, in the younger or older terraces, seemingly makes testing of any postulated uplift model difficult.

For the younger (post-Rapanui) terraces, simply on a count-forward-from-120Ka BP basis, the Inaha Terrace might be expected to have been cut during the widely recognised interstadial high sea level at c.100Ka BP (Table 5.6). Similarly, the unnamed post-Inaha terrace might be expected to have been cut at c.80Ka BP (Table 5.6). Sea level during both these transgressions apparently reached c.-15m relative to present (Table 5.6). To test these assumptions, two uplift models for the younger South Taranaki terraces are considered:
(1) uniform uplift rates, equal to those of the Rapanui strandline

(2) hinging-type uplift, equal to the tilt rate between the Rapanui and Kaiatea III strandlines.

Results are given in Table 5.8. I conclude that model 2 yields paleosealevel estimates in closest agreement with those from overseas. Ages of c.100 and 80Ka BP for the Inaha and post-Inaha strandlines respectively are therefore accepted.

A further check on the above models is provided by the small terrace remnant west of the Whenuakura River (see p.111), which may represent a yet younger terrace than the post-Inaha terrace. An interstadial high sea level stand at c.-28m relative to present is dated in New Guinea at c.60Ka BP (Bloom et al. 1974) and is dated in Barbados (James et al. 1971), Japan (Machida 1975) and the Ryukyu Islands (Konishi et al. 1970). Assuming an age of 60Ka BP for the South Taranaki terrace remnant, model 1 yields a paleosealevel estimate of -30m, while model 2 yields an estimate of -24m relative to present (Table 5.8). In other words either model yields paleosealevel estimates in accord with the New Guinea data. An age of 60Ka BP is therefore adopted for this terrace remnant.

For the older (pre-Kaiatea III) terraces, the same two models are again considered. In this case, however, little control is available using overseas data because of the lack of paleosealevel estimates. Consequently, it must be assumed that all sea levels associated with the pre-Kaiatea III strandlines were similar to present. Based on strandline heights given in Table 5.2, age estimates are made for the Kaiatea II and I strandlines, as well as the two possible pre-Kaiatea strandlines; these are listed in Table 5.9 for both the hinging-type and uniform uplift models. The hinging-type uplift model yields ages which tend to cluster between 400 and 500Ka BP, which seems unreasonable in the light of deep sea core evidence.
and overseas sea level data (Table 5.6, Fig. 5.4). The uniform uplift model yields age estimates which are in moderate agreement with major high sea level events identified elsewhere. The uniform uplift model is therefore tentatively adopted. If the actual pattern of deformation includes a small tilt component, the derived strandline ages based on this uniform uplift model are minimum ages.

To summarise, the deformation pattern in South Taranaki is approximated by

(1) constant uplift inland of the Kaiatea III strandline
(2) tilting uplift seaward of the Kaiatea-III strandline.

Uplift rates, are shown in Fig. 5.5 and vary between 0.66mm/year and 0.29mm/year.

To estimate final model ages of individual terrace strandlines, a relation diagram is constructed based on uplift rates given in Fig. 5.5, and strandline heights in Table 5.2; the relation diagram is shown in Fig. 5.6. For all strandlines older than the Rapanui (c.120Ka BP), sea level is assumed similar to the present. Best fit linear regressions of the form

\[ Y = AX \]

where \( Y \) = strandline height (m)
\( A \) = slope = age (Ka)
\( X \) = uplift (mm/year)

are given for each strandline. Strandline height data for the Ngarino Terrace (Table 5.2, Fig. 5.1) suggest that the Ngarino terrace might be a composite feature. On this basis, two possible age groups are shown in the relation diagram for the Ngarino strandline.
For strandlines younger than the Rapanui, linear regressions of the form

\[ Y = C + AX \]

where \( Y \) = strandline height (m)
\( C \) = paleosealevel relative to present (m)
\( X \) = uplift rate (mm/year)
\( A \) = age (Ka)

have been fitted to the data.

Resultant best estimate ages (based on deformation models, fission track and amino acid racemisation dates, as well as overseas data) for all strandlines are summarised in Table 5.10. Strandline ages are also shown in Fig. 5.4 for comparison with overseas data.

5.4 RELATIONSHIPS BETWEEN THE TERRACES AND YOUNGER MARINE SEDIMENTS IN THE WANGANUI BASIN

The Landguard Formation at Landguard Bluff near the mouth of the Wanganui River is the youngest formation of the Castlecliffian Stage, with all younger sediments being attributed to the Hawera Series (Finlay & Marwick 1947, Fleming 1953, Suggate 1962). Traditionally, all terraces of the South Taranaki-Wanganui sequence have been assigned to the Hawera Series, although previously only the Rapanui Terrace has been demonstrated by superposition to be younger than the Landguard Formation. Several workers (eg. Fleming 1953, Grant-Taylor & Te Punga 1978) have remarked that the Kaiatea terraces, at least in part, could belong to the Castlecliffian Stage. However, except for the assertion of Grant-Taylor & Te Punga (1978, p.552) that "the plane
of the oldest Kaiatea bench projects seaward into the Shakespeare Group..." and their inference that Kaiatea sediments may have originally been conformable on the Shakespeare Group towards the coast, no firm evidence of the relationship has ever been forthcoming.

As indicated previously (p.113), part of the reasoning for including all terrace deposits in the Hawera Series lies with the fact that the Hawera Series/Castlecliffian Stage boundary has been seen to represent a regional unconformity, (see Fleming 1953, p.247 for example), separating deposits clearly related to the present landscape (Upper Quaternary) form underlying deposits which are in most instances quite discordant with present topography (Lower Quaternary), and more closely similar to those of the Pliocene. However, more than 20 years ago Te Punga (1957a) argued that this postulate was not tenable on a regional basis, when he described a conformable sequence of sediments in the Rangitikei Valley east of Wanganui, which passed upwards from the Mingaroa Fossil Beds (correlated with the Landguard Formation by Fleming (1953, 1957a) to sediments of Hawera Series age. In other areas, relationships between the Hawera Series and the Castlecliffian Stage are hindered by lack of stratigraphic and chronologic control, so that the existence of a regional break between the two must be regarded with some suspicion. At best, the proposition of discordance/accordance with present landscapes is probably best viewed as a convenient simplification, applicable towards the Hawera-Wanganui Coast, where younger terraces of the Hawera Series are clearly cut into older Wanganui Series beds.

Work in this study has shed considerable light on Hawera Series/Castlecliffian Stage relationships as follows:
(1) The Brunswick Formation has been demonstrated to stratigraphically overlie the Landguard Formation, perhaps conformably. This is based on a section near Warrengate Trig, east of Wanganui (N138/732815; see Fig. 4.6), where a shellbed containing *Pecten n. aotea*, an index fossil of the Landguard Formation, underlies the Brunswick Terrace. This relationship is also supported by the 310Ka BP age estimate for the Brunswick strandline (Table 5.10) and the 340Ka BP age estimate by Boellstorff & Te Punga (1977) for the top of the Castlecliffian Stage.

(2) The Rangitawa Pumice occurs within uppermost Castlecliffian marine sediments at its type section (N143/962624; Fig. 4.1). If the correlation of the Rangitawa Pumice with the Omahina Tephra is accepted (Pillans & Kohn, in press), the Kaiatea III Formation must be of Castlecliffian age and is therefore the nearshore equivalent of some part of the Upper Castlecliffian Wanganui Basin sequence. Even if the correlation of the Rangitawa Pumice is rejected, the fission track dates still support a Castlecliffian age for the Kaiatea III Formation (c.370Ka age for the Omahina Tephra, and c.340Ka age for the top of the Castlecliffian Stage).

(3) The age of the Upper Brunswick strandline is estimated to be 340Ka BP (Table 5.10). This age is identical with the age estimate of 340Ka BP made by Boellstorff & Te Punga (1977) for the top of the Castlecliffian Stage. On this basis, a bored waved cut surface, which lies c.3m above the shellbed containing *P.n.aotea* and below the interpreted base of the Brunswick Formation at the Warrengate section, is correlated with the Upper Brunswick wave cut surface. The age of the
Upper Brunswick Formation is therefore inferred to be close to that of the Hawera Series/Castlecliffian Stage boundary.

If it is assumed that the deeper water formations of the Castlecliffian Wanganui Basin sediments accumulated during times of major sea level rise (cf. Vella (1963) with respect to Lower Pleistocene Wanganui Series beds), then these are the most logical correlatives of the shallow marine Kaiatea and pre-Kaiatea terrace deposits. Fission track dates from tephras within the Wanganui Basin sequence (Seward 1974, 1976, 1979) provide chronologic control which allows reasonable correlations to be made with the South Taranaki terrace sequence. An independent check on fission track dates is possible because the Brunhes/Matuyama paleomagnetic boundary is located between the Potaka Pumice and the Rewa Pumice (both of which have been fission dated) at the base of the Kai-Iwi Group (Seward 1974). Since age estimates for the South Taranaki Terraces (Table 5.10) are less than the age of the Brunhes/Matuyama boundary (c.730 Ka BP; Mankinen & Dalrymple 1979), the base of the Kai-Iwi Group forms a convenient lower boundary for the correlations described here. The stratigraphic sequence between the base of the Kai-Iwi Group and the top of the overlying Shakespeare Group (top of the Castlecliffian Stage) in the Wanganui area is shown in Table 5.11 (after Fleming 1953). Correlations with the Rangitikei Valley sequence (Te Punga 1952) which follow those of Fleming (1953, 1957a), are also shown in Table 5.11. Fission track dates are also shown.

Fleming (1953, 1978) has indicated that the Pinnacle Sand and Tainui Shellbed are the deepest-water formations of the Shakespeare Group. Furthermore, "they contain warm water mollusca of subtropical origin (Pterynotus zealandicus, Zelippistes) and pollen of the Kauri (Agathis), now confined to Northland" (Fleming 1978, p.563). Seward
(1974) has dated the overlying Rangitawa Pumice at 380±40Ka BP and the underlying Waiomio Shellbed at 450±90Ka, so that correlation of these two formations with the cutting of the Kaiatea III wave cut surface (at c.400Ka BP) is possible (see Table 5.11).

In the Wanganui region, the Kupe Formation contains several species now confined to northern New Zealand, and Fleming (1953, p.205) suggested that part at least was deposited in subtropical waters 3-6°F warmer than those of Cook Strait today. The overlying Upper Kai-Iwi Siltstone is the deepest water formation of the Kai-Iwi Group (Fleming 1953, Fig. 62, p.302). The Waiomio Shellbed dated at 450±90Ka BP is correlative with the Kupe Formation, so that correlation of the Kupe Formation/Upper Kai-Iwi Siltstone with the cutting of the Kaiatea II wave cut surface (at c.440Ka BP) is possible.

On the basis of the above correlations, deposition of the Seafield Sand at the base of the Shakespeare Group must have occurred between the cutting of the Kaiatea II and III wave cut surfaces. Fleming (1953, p.113) reported that on the coast west of Wanganui (near N137/471900), the base of the Seafield Sand unconformably overlies the Upper Kai-Iwi siltstone, the top of which is bored by Anchomasa similis; this he interpreted as a wave cut surface. If the transgression following the cutting of this surface culminated at the Kaiatea III strandline, an estimate of the relative rise in sea level is possible using the uplift rates summarised in Fig. 5.5. On traverse 13 (Fig. 5.5), the mean uplift rate is c.0.28mm/year at the coast (where the Seafield Sand is exposed), and is c.0.52mm/year at the Kaiatea III strandline. Assuming an age of c.420Ka BP for the Seafield Sand/Kaiatea III transgression a constant rate of shore normal tilting since that time would produce a height difference of c.100m between the base of the Seafield Sand on the coast and the
Kaiatea III strandline. Since the present height of the Kaiatea III strandline on traverse 13 is c.206m (Table 5.2) and the base of the Seafield Sand is near present sea level, a eustatic rise of c.100m is inferred for the Seafield Sand/Kaiatea III transgression.

[NOTE: That if the transgression which followed the deposition of the Seafield Sand culminated at the Kaiatea II strandline a eustatic rise of c.100m is also indicated (ie. if both the Kaiatea III and Kaiatea II strandlines correlate with the Tainui Shellbed/Pinnacle Sand deep water). This possibility seems less likely at this stage, but cannot be entirely ruled out given the imprecision in the fission track age of the Kupe Formation].

Correlation of the Kaiatea I Terrace (c.520Ka BP) is more difficult because the stratigraphic record near Wanganui becomes compressed in the lower portion of the Kai-Iwi Group. The Kai-Iwi Group thickens eastwards, but the critical part of the basin in the Whangaehu Valley has yet to be studied in detail. The Waitapu shellbed in the Rangitikei Valley further east has been dated at 520±80Ka BP by Seward (1974), and is correlated with the Omapu shellbed near Wanganui (Fleming 1953). Immediately underlying the Omapu shellbed is the deeper water Lower Kai-Iwi Siltstone which contains the Kaukatea Ash dated at 570±80Ka BP (Seward 1976). Tentative correlation of the Kaiatea I wave cut surface is made with the younger part of the Lower Kai-Iwi Siltstone.

The two older pre-Kaiatea I terraces (estimated to be 600 and 680Ka BP respectively) appear to correlate approximately with the Kaimatira Pumice Sand at the base of the Kai-Iwi Group. The older of these two terraces should contain the Potaka Pumice (dated at 610±60Ka (Seward 1976) and 640±180Ka BP (Seward 1979)), if the fission track
age estimates are close to the real age of the Potaka Pumice and if the age estimate made in this work for this older terrace is correct. Further study of this possibility is recommended, as the presence of a tephra older than the Omahina Tephra within the terrace cover beds would be extremely valuable to the terrace chronology.

5.5 RELATIONSHIPS BETWEEN THE SOUTH TARANAKI TERRACES AND VOLCANIC DEPOSITS IN WESTERN TARANAKI

According to Grant-Taylor (1964a, b) Quaternary marine terrace formation in western Taranaki has alternated with aggradation of breccias and conglomerates produced by lahars (volcanic mudflows) originating from the Egmont, Pouakai and Kaitake volcanic centres. These lahars, as well as other volcanic lastics, form a gently dipping surface around the lower slopes of Mt Egmont; this surface, and its associated lithologies, have been termed a volcanic ring plain (Morgan & Gibson 1927). An older, much more dissected ring plain, which is thickly mantled with tephras, is also present around Pouakai. Only isolated remnant ridges remain of the older Kaitake ring plain (Neall 1979).

Grant-Taylor (1964a, b) recognised and named seven major groups of lahars in western Taranaki (Table 5.12). Pollen assemblages from the Opunake and Stratford Lahars (Egmont Source), Maitahi Lahars (Pouakai source) and the Eltham Lahars (Kaitake source) all indicated cold climates equivalent to those of 1000 to 1500m on Mt Egmont today (Grant-Taylor 1964a, b). No pollen data was obtained from the Lepperton Lahars (Egmont and/or Pouakai source) and the Inglewood and New Plymouth Lahars (Kaitake source); the absence of warm climate florals in those deposits examined, led Grant-Taylor to conclude that the periods when the ring plain lahars accumulated were
glacial stages. However, more recent work by Neall (1972, 1979) has shown that several major lahars have been produced on Egmont in the last 10,000 years, so that not all lahars can be inferred to be of glacial climate origin in Taranaki.

Marine terraces in western Taranaki, and correlated with the Rapanui, Ngarino and Brunswick Terraces of South Taranaki were mapped by Hay (1967), who also adopted the same names for their supposed western Taranaki correlatives. Stratigraphic relationships between the lahars and these marine terraces as shown by Hay (1967), follow those suggested by Grant-Taylor (1964a, b) - see Table 5.12. Stratigraphic relationships between supposed correlatives of the Kaiatea I, II and III terraces in South Taranaki, as summarised by Grant-Taylor (1964a, b) and Dickson et al. (1974), are also given in Table 5.12. Unfortunately the lahar units and marine terraces recognised by Grant-Taylor (1964a, b) have never been stratigraphically adequately defined: no type sections have ever been named. Consequently the stratigraphy erected by Grant-Taylor has never been properly described or adequately tested. Neall (1979) remapped a large portion of the volcanic deposits in western Taranaki and failed to recognise both the New Plymouth and Inglewood Lahars of Grant-Taylor (1964a, b) and Hay (1967); further field work is warranted to more fully test this possibility.

Unpublished K/Ar dates (Stipp 1968) and published $^{14}C$ dates (Grant-Taylor 1964a, b, 1978b; Neall 1979) allow ages of the lahar deposits to be estimated (Table 5.12). Stratigraphic relationships and ages determined in this study, and those of previous workers (summarised by Neall 1979) permit better correlation than was previously possible between the lahars and the South Taranaki marine terraces. In addition, qualitative assessment of the variation in
supply of volcanic detritus to the South Taranaki terrace sequence supports the correlations made here (see Table 5.12). Detailed study of the provenance of marine sands within the South Taranaki terraces has not been attempted in this study, but future work of this kind would be most instructive.

At the Inaha type section (N129/748278) a $^{14}$C date of 33,000±1100 years BP (ANU-1887) was obtained in this study (Table 3.3; Fig. 4.10) from a lignite interbedded with laharic debris in the upper portion of the section. Grant-Taylor & Rafter (1963) reported dates of 34,400±1500 years BP (NZ331A) and 31,800±1800 years BP (NZ409A) from the Opunake Lahars some 30Km west of the Inaha type section, while Neall (1979) has mapped the Opunake Formation eastwards from there to within 1Km west of the Inaha type section. Correlation of the lignite at the Inaha section with the Opunake Lahars (Opunake Formation) is thus strongly supported. The Opunake Formation therefore stratigraphically overlies the Inaha Formation at this locality; however, until the base of the Opunake Formation is fixed, further comment is not possible.

According to Grant-Taylor (1978b), in North Taranaki, near Waitara (some 30Km northeast of New Plymouth), the Opunake Formation directly overlies a lignite containing tree stumps in growth position. Pollen samples from this lignite (referred to as the Waitara Formation) indicated a forest dominated by Dacridium cupressinum (rimu) and Metrosideros robusta (rata) (Grant-Taylor 1978b). A $^{14}$C date of >50,000 years BP (NZ406) from the Waitara Formation, and a $^{14}$C date of 34200±1500 years BP (NZ407) from the Opunake Formation at this site were also reported by Grant-Taylor (1978b). The next oldest group of lahars, the Stratford Formation (Grant-Taylor & Kear (1970), Neall (1979)), apparently directly underlies the Waitara
Formation (Grant-Taylor 1978b). At the Inaha type section (N129/748278) pollen samples from the Manaia Lignite (McGlone et al 1979) record a forest phase between the Inaha Formation (c.100KA BP) and the uppermost lignite (c.33 Ka BP); all other floras at the Inaha section are grass/scrub dominated (McGlone et al 1979). If correlation of the Manaia Lignite with the Waitara Formation is accepted, the Stratford Formation must predate the Inaha Formation and cutting of the Inaha wave cut surface.

On the coast near Hawera, lahar deposits which are correlated with the Stratford Formation by Grant-Taylor (1978b) stratigraphically overlie the Rapanui Formation. Once again the lack of stratigraphic definition precludes closer analysis of relationships between the two; however, since the Stratford Formation is younger than the Rapanui Formation, and likely to be older than the Inaha Formation, an age of between 100 and 120Ka BP is indicated.

Stratigraphic relationships as far as they can be determined between the South Taranaki terraces and the volcanic deposits in western Taranaki are summarised in Table 5.12.

5.6 NEW ZEALAND QUATERNARY CHRONOLOGY

The currently accepted New Zealand Quaternary chronology is essentially that of Suggate (1960, 1962, 1965, 1978a), in which a number of climatically defined stages have been established based on outwash surfaces and moraines in the South Island (glacial) and uplifted marine terraces in both the North and South Islands (interglacial) - see Table 5.13 for summary.
According to Suggate (1965, p.76):
"the main critical points of climatic change are those of the commencement of rapid cooling leading to glaciation and of rapid warming leading to interglacials". He therefore defined and named the seven stages listed in Table 5.7 in these terms. For example, (Suggate 1965, p.83), the Oturian Stage was defined as
"embracing all those deposits laid down between the beginning of the warming that led to the Oturi Interglacial and the beginning of the cooling that led to the Otira Glaciation"
Similarly the Waimean Stage was defined as embracing all those deposits laid down between the beginning of the cooling that led to the Waimea Glaciation and before the warming that led to the Oturi Interglacial.

However, as Carter (1974) has pointed out, these definitions are unacceptable for time stratigraphic/stage units because they:
(1) assume a priori that the pattern of warming and cooling was simple,
(2) require the location of two assumed points of maximum climatic change, which themselves are time transgressive.
Furthermore he argued that if the Quaternary "Stages" were to form part of a local time scale, they must be adequately defined:
"preferably by 'golden pegs' in type sections; otherwise their 'i'an ending and the class-noun 'stage' should be dropped" (Carter 1974, p.194).

The situation is further complicated by the fact that the Terangian and Oturian Stages were originally defined at type sections in the South Taranaki-Wanganui terrace sequence, but were subsequently redefined by Suggate (1965) to represent the penultimate and last interglacials respectively. Such a redefinition appears to have been made largely on a countback from present basis: the Rapanui (and Ngarino terrace) was the first major terrace and the Brunswick was the second major terrace inland from the present coastline. Consequently the redefinition was acceptable as long as the Brunswick and Rapanui
Terraces could be correlated with the penultimate and last interglacials respectively (or as Suggate (1965, p.81) stated "unless and until conditions are proved cold enough to indicate a glacial period between the two younger (Ngarino and Rapanui) cliffs at Wanganui."). This meant that the Terangian type section was still correlated with the penultimate interglacial and the Oturian type section still correlated with the last interglacial. Milne (1973a) apparently confirmed that such a correlation was in accord with radiometric dates from New Guinea and Barbados, ie.

- Rapanui Terrace = c.120Ka (penultimate) interglacial
- Brunswick Terrace = c.200Ka (penultimate) interglacial.

However, the implications of dates presented in this work are that

1. the Brunswick Terrace, and hence the type locality of the Terangian Stage, cannot be correlated with the penultimate interglacial (Termination III),
2. the Rapanui Terrace can be correlated with the last interglacial (Termination II), but that the type locality of the Oturian Stage does not occur on the Rapanui Terrace.

Consequently there appear to be discrepancies in the existing New Zealand chronology of stages in the Upper Quaternary. Possible modifications of the present chronology are now discussed in the light of South Taranaki data.

Table 5.14 summarises present New Zealand Quaternary chronology with respect to the South Taranaki Terraces and two possible modifications.

On the basis of inferred stratigraphic relationships between the Mt Curl Tephra (see Appendix 3), and the Brunswick Terrace east of Wanganui, Milne (1973a, b) estimated that the Terangi ("penultimate") interglacial commenced at 230Ka BP ie. the Brunswick wave cut surface began forming at about this time. However, using the chronology developed in this work, the age of the Brunswick wave cut surface is estimated to be 310Ka BP; in other words, the type
formation of the Terangian Stage (Terangi Interglacial) is apparently considerably older than previously estimated.

Accepting the chronology presented in this study, and following previous correlations of the Brunswick wave cut surface with the Terangian Interglacial and the Rapanui and Ngarino wave cut surfaces with the Oturi Interglacial (cf. Suggate 1965), a possible revision of New Zealand Upper Quaternary Stage boundaries would be as shown in Fig. 5.7 column B (see also Table 5.14). According to this scheme, the Mt Curl Tephra would lie close to the boundary of the Oturian and Waimean Stages, while the Terangian/Waimaungian Stage boundary would closely approximate the Hawera Series/Castlecliffian Stage boundary.

If, on the other hand, the Terangi Interglacial is equated with the penultimate interglacial (Termination III of Broeker & Van Donk (1970)) and the Oturi Interglacial with the last interglacial (Termination II), the New Zealand Upper Quaternary Stage boundaries would be as shown in Fig. 5.7 column D (see also Table 5.14). According to this scheme, the Mt Curl Tephra would be close to the Terangian/Waimaungian Stage boundary and the lower boundary of the Waiwheran Stage would closely approximate the Hawera/Castlecliffian boundary. Furthermore the present type locality of the Terangian Stage would become the type locality of the Waiwheran Stage.

Of the above two alternatives, the latter is preferable for several reasons:

(1) The recognition of a glacial event between the formation of the Rapanui and Ngarino wave cut surfaces appears warranted in view of the suggested correlation with the oxygen isotope record in core V28-238 (Fig. 5.7).

(2) The placing of the stage boundaries approximates those ages adopted by Milne (1973a, b) for the Terangian and younger stages. These in turn he subdivided into named substages based on loess stratigraphy. Thus the loess chronology would remain unchanged except for its inferred relationships with the South Taranaki marine terraces.
(3) The number and timing of named New Zealand Upper Quaternary stages would closely match those oxygen isotope stages recognised in Pacific core V29-238 by Shackleton & Opdyke (1973) and which has been recommended by them as a standard Quaternary chronology. Only stage 3 of Shackleton & Opdyke would not have a named equivalent in New Zealand, although the c.60Ka marine terrace can probably be correlated with this stage.

Accepting this second alternative (Fig. 5.7 column D), the following stage boundaries could be defined in South Taranaki-Wanganui:

(1) Base of the Otiran Stage: top of the Manaia Lignite* in the Inaha type section (Fig. 4.10) at N129/748278
(2) Base of the Oturian Stage: Rapanui wave cut surface beneath the Denby Shellbed (Fig. 4.8) at N129/820259
(3) Base of the Waimean Stage: top of esturine sediments at the new Ngarino type section (Fig. 4.7) at N130/217108
(4) Base of the Terangian Stage: Mt Curl Tephra (Appendix 3)
(5) Base of the Waimaungan Stage: top of estuarine sediments at the Brunswick type section (Figs. 4.4, 4.5) at N137/502981
(6) Base of the Waiwheran Stage: Upper Brunswick wave cut surface at the Warrengate Section (Fig. 4.6) at N138/732815, which also approximates the Hawera/Castlecliffian boundary.

This would therefore satisfy the requirement of Carter (1974) that stage boundaries be defined by "golden pegs" in type sections.

* The Manaia Lignite contains a flora dominated by Podocarpus dacrydioides (Kahikatea or white pine) and Podocarpus spicatus (matai), which according to McGlone et al (1979) indicates a climate neither as warm or as wet as present, and suggests that the Manaia Lignite formed during a warm interstadial event post-dating the cutting of the Inaha wave cut surface (c. 100Ka BP); correlation of the Manaia Lignite with the c.80Ka BP interstadial high sea level (post-Inaha wave cut surface) seems likely.
CHAPTER 6

GEOMORPHOLOGY

6.1 INTRODUCTION

As indicated in Chapter 1, the primary aim of this study is to quantify the long term geomorphic evolution of the South Taranaki terrace sequence. Towards this end, considerable time has necessarily been devoted to dating the terraces, describing their stratigraphy, and reconstructing their geological history. This geological data will now be used, in conjunction with geomorphic data described and analysed in this chapter, to form the basis for quantitative evolutionary models which are presented in Chapter 7. In this respect, the terraces themselves are to be used as dated surfaces, each similar to the next except for age differences: the landforms developed on the terraces are therefore inferred to represent an evolutionary sequence, whose initial forms may be reconstructed and whose rates of development may be quantitatively assessed.

Even at first sight, it may easily be discerned that the extent of terrace dissection by fluvial processes increases up the terrace flight; valleys become wider and deeper, drainage densities increase, until eventually whole terrace surfaces are completely dissected (Plate 6.1). In other words the "geomorphic maturity" increases up the terrace flight. These rather simple notions of fluvial evolution are quantified in Fig. 6.1, which shows a plot of (A) percent interfluve area and (B) drainage density versus terrace age for the Rapanui, Ngarino, Brunswick and Kaiatea III terrace surfaces in the Waverley district. Best fit linear regressions indicate that drainage density approximately doubles every 100Ka, while 50% of terrace surface is dissected in c.260Ka. If this linear trend of terrace dissection is extrapolated, a terrace surface would
be completely dissected in slightly more the 500Ka - this is in accord with the observation that only minor remnants of pre-Kaiatea III terraces are present inland from Waverley (see Map 1).

Following emergence of a particular marine terrace, terrestrial sediments (eg. tephra and dunesand in particular) slowly accumulate on its surface. This, in general, leads to progressive burial of the initial groundsurface and results in increasing non-marine cover bed thickness with time. This trend is shown in Fig. 6.1C which is a plot of mean estimates of non-marine cover bed thickness measurements versus terrace age for the post-Inaha, Inaha, Rapanui, Ngarino, Brunswick, Upper Brunswick and Kaiatea III terraces in the Waverley area. A best fit linear regression indicates that 1 metre of non-marine cover beds accumulates every c.16Ka or 6m/100Ka, with an intercept of zero non-marine cover at age zero. Progressive fluvial dissection of terrace surfaces is therefore accompanied by progressive accretion in interfluve areas.

Immediately following emergence of a terrace surface, shallow marine and littoral sediments would be virtually the only deposits overlying the wave cut surface. The initial groundsurface can therefore be approximated to the top of the marine sediments. This initial surface is assumed to be a gently sloping, planar surface corresponding to the marine/non-marine interface within any given terrace, as field observations do not allow mere exact definition. Owing to progressive accretion of non-marine cover beds, interfluve areas therefore preserve a buried record of this initial form.

The time of initiation of fluvial dissection on each terrace must follow shortly after emergence of the initial surface. Since the regional dip of this initial groundsurface would have been seawards, dominantly shore-normal drainage would be expected in the early stages
of landscape development (Fig. 6.2A). Streams developed in this way are referred to as "consequent" streams in classical geomorphological nomenclature eg. Cotton (1942, p.42). Since the age of a particular valley can therefore be no older than the terrace surface it dissects, and if its initiation is assumed to have occurred at emergence, the terrace strandline ages given in Table 5.10 represent close approximations to time of initiation of fluvial dissection on their respective terrace surfaces.

Several types of fluvial landforms may be distinguished on the terrace sequence as follows:

(1) Major shore-normal river valleys which traverse the entire terrace sequence and originate inland of the terraces (Plate 2.6, Fig. 2.1).

(2) Small valleys and drainage nets incised into terrace surfaces and draining directly into major rivers; largely shore parallel (Plate 6.3).

(3) Larger tributaries and drainage nets cut into terrace surfaces, sometimes traversing several terrace surfaces.

(4) Small valleys and drainage nets confined to single terrace surfaces, and generally aligned in a shore-normal direction, of which two classes are recognised:

(a) Those confined to terrace margins (single small valleys predominate) eg. Plates 2.10 and 6.9.

(b) Those which traverse broad terrace surfaces (drainage nets predominate) eg. Plate 6.1.

(5) Ponded stream valleys inland of encroaching coastal dunes.

The development of these fluvial systems has not occurred on a homogeneous, uniformly dipping initial surface as was envisaged by Horton (1945, p.343) when he qualitatively described the origin and development of streams on a coastline with falling relative sea level (or as described above for individual terrace surfaces). Rather, the history of fluvial dissection in South Taranaki is strongly influenced by subsequent cliffing of terrace outer edges by later marine transgressions (Fig. 6.2B). The cliff produced during and up to the culmination of these subsequent transgressions affects the course of
fluvial dissection not only by simply truncating older emergent surfaces and fluvial systems, but also by causing an increase in local relief along which renewed valley initiation and/or dissection may occur as follows:

Consider firstly the case of a small shore-normal valley being truncated by the retreating coastal cliff, in which the rate of cliff retreat far exceeds the rate of stream downcutting; continued truncation of the upstream valley segment occurs, and a "hanging-valley" situation is maintained at the receding cliff front (Fig. 6.2C). Such a situation is likely to occur while relative sea level is rising and the rate of cliff retreat high. Secondly, consider the case where the rate of cliff retreat is roughly equal to, or exceeded, by the rate of stream downcutting, eg. stable relative sea level at culmination of transgression with low rate of cliff retreat (Fig. 6.20). In this second case, stream downcutting at the cliff front results in the development of a knickpoint separating a downstream valley segment, with steep gradient (including waterfalls and rapids perhaps) from an upstream valley segment which is little changed from its pre-transgression form (Fig. 6.20).

Observations along the present coastal cliffs, which are retreating at a mean rate of c.70cm/year (Gibb 1978), provide insights into these early stages of terrace-front valley evolution, and confirm the general evolutionary sequence outlined above: smallest streams discharge into the sea via waterfalls and/or steep gorges near the present cliff front; an upstream valley segment is present above a well defined knickpoint and is presumably representative of the pre-cliffing, terrace-surface, drainage system. Larger streams, however, lack these features indicating a different evolutionary sequence accompanying cliffing (Figs. 6E,F).
In the case of larger shore-normal valleys, the rate of downcutting is such that the rate of incision is always likely to be roughly equal to or greater than the rate of cliff retreat. Consequently, the "hanging-valley" form is never evident. Postulated evolutionary forms are as shown in Figs. 6E and 6F.

Thus, on any particular terrace surface, it is possible to recognise two major stages of valley development, particularly among the smallest terrace front drainage networks (see 4a, b above):

Stage 1: Initiation of terrace surface drainage (consequent streams) immediately following emergence (Fig. 6.2A).

Stage 2: Truncation of terrace surface drainage and incision near the terrace front (evident along the present coastline) during the following marine transgression (Figs. 6.2C, D).

In addition a further stage may be recognised on older inland cliffs, but which has not yet occurred on the present coastal cliffs -

Stage 3: Valley initiation along the terrace front, following abandonment of the cliff during the subsequent regression. (Presumably the present rate of active sea-cliff retreat is greater than the headward rate of growth of these incipient terrace front valleys).

The above discussion highlights the problem of assigning ages to particular valleys on the South Taranaki terraces, and may be summarised as follows: Valleys must be younger than the terrace surface they dissect, but not appreciably younger if valley initiation takes place soon after emergence (Stage 1 above). Unfortunately valleys may also be appreciably younger than the terrace surface they dissect owing to more recent cliffing, although once again valley ages may be estimated by assuming that valley initiation along the cliff front occurred soon after abandonment of the cliff (Stage 3 above).

However, while the smallest terrace front valleys (case 4(a) above) might reasonably be inferred to post-date the cliff, slightly larger stream valleys, perhaps crossing an appreciable part of a terrace surface, may not necessarily relate to this later event.
It is appropriate at this point to introduce two major schools of thought regarding valley development on initial surfaces. The first of these is the headward retreat school, in which drainage networks are viewed as expanding with time by headward extension and branching processes (in much the same way as tree growth). This approach was adopted by Horton (1945), in which he postulated that the extension process was limited largely by the critical distance of overland flow (or belt of no erosion), and was also adopted by Dunkerley (1977b) who argued that extension was governed by the constant of channel maintenance i.e. drainage network expansion did not take place unless a certain critical distance (Horton) or critical area (Dunkerley) was exceeded. In this case therefore, the time of initiation of a particular valley or valley segment might conceivably be at any time following terrace emergence. Faced with this concept, Carter & Chorley (1961, p.128) argued that "a stream segment of one order may be expected to be constantly and progressively changed into one of higher orders as channel extension...occurs to the headward of it". They therefore concluded that valley age was proportional to stream order in the expanding stream network they studied.

The second major view of valley development is essentially that of valley deepening and widening as exemplified by W.M. Davis's cycle of erosion, and more recently by Chappell (1974a) in his modelling of stream thalweg evolution on an Upper Quaternary flight of coral terraces in New Guinea. Mizutani (1974) also followed this approach in his modelling of the dissection of volcanoes in Japan. In these cases, the headward extent of valley systems is seen as being fixed, and evolution proceeds largely through thalweg lowering.

The New Guinea coral terraces studied by Chappell (1974a) and Dunkerley (1977a, b) provide an interesting comparison with the South
Taranaki terraces described in this study. However, the New Guinea terraces are largely features constructed by coral reef growth, where subsequent transgressions and terrace building episodes cause negligible cliff retreat, and accompanying incision/valley initiation effects differ from the South Taranaki terraces. In this way Chappell (1974a) was able to select small terrace front valleys which were initiated shortly after terrace emergence ie. their age could be assumed to closely post-date the age of emergence of the reef crest which they dissected.

With the above considerations in mind, this chapter seeks to understand and quantify the evolution of a suite of simple fluvial landforms on several of the South Taranaki terraces: small terrace front valleys, which are inferred to have been initiated shortly after cliff retreat ceased (Stage 3 and category 4(a) above), and analogous in some respects to those described by Chappell (1974a) in New Guinea.

6.2 VALLEY INITIATION BY TERRACE FRONT STREAMS

A striking feature of the South Taranaki terrace sequence is the difference in properties of the terrace cover beds (both marine and non-marine) as compared with those of the underlying Tertiary-Lower Quaternary sediments. For example, the cover beds tend to be

(a) less consolidated
(b) more porous
(c) more weathered

than the underlying strata, particularly in the western half of the terrace sequence, where Tertiary mudstones and siltstones are predominant. This means that wave cut surfaces are geomorphically significant boundaries (Plate 6.2).
To illustrate, subsurface water apparently moves through the cover beds with relative ease, but is prevented from penetrating the underlying Tertiary strata to any great extent. As a result, in many places where wave cut surfaces intersect the present ground surface (eg. cliffs, valley sides), groundwater seeps out just above the wave cut surface and trickles down the face of the underlying Tertiary exposure. Wave cut surfaces are therefore marked by vegetation lines, which are particularly noticeable in areas cleared of natural forest (Plates 6.5, 6.6, 6.11). Subsurface flow must be preferentially concentrated into seepage lines or subsurface streams because discharge along wave cut surfaces is not uniformly distributed; rather the discharge is via springs (point sources), some of which have discharges in excess of several litres/minute. These springs have been observed in the modern coastal cliff, along valley sides and in valley heads, and many provide a permanent source of domestic and stock water.

Subsurface flow is strongly controlled by the seaward dip of wave cut surfaces (see 5.1 Terrace deformation: principles), despite the fact that these dips are mostly less than 1° from horizontal (Table 5.1). This is clearly seen in Plates 6.3 and 6.4, where a small shore-parallel valley is cut approximately at right angles to the dip of the wave cut surface (which dips from left to right in Plate 6.3). On the updip side of the valley (left side) many small springs are present along the intersection of the wave cut surface with the present ground surface; on the downdip side no springs are present. This phenomenon is widely exhibited by shore-parallel valleys (ie. normal to wave cut surface dips) on the terraces, and clearly indicates a strong component of shore-normal subsurface flow (down dip). Water from each spring (Plates 6.3 and 6.4) flows down a shallow, swampy channel towards the valley floor.
This results in well developed spur and groove topography on the updip valley side, while the downdip side remains comparatively smooth (Plate 6.3). Also note that the downdip valley-side slope is steeper than the updip valley-side slope. Similar spur and groove topography is also present below wave cut surface springs on valley sides of shore-normal streams (Plates 6.5 and 6.6).

Immediately upslope of these springs are small hollows with steep, terracedding headwalls and swampery, gently sloping floors (Plates 6.5 and 6.6). These hollows, or seepage heads, are developed entirely within the cover beds and appear to be formed by slumping which is caused by increased soil moisture immediately adjacent to the springs (also accentuated by the unconsolidated nature of the cover beds). Once formed, the seepage heads tend to concentrate surface and subsurface water which not only promotes further slumping, but also facilitates removal of slumped material. This positive feedback situation leads to general enlargement of the hollow, and, in particular, headward retreat in the direction from which most seepage is received. Large amphitheatre shaped slump heads within the terrace cover beds are the result (Plate 6.5).

Selby (1966, 1967) has described similar amphitheatres from elsewhere in the North Island of New Zealand and suggested that such features resulted from slumping within weathered ash and greywacke overlying unweathered greywacke. He also noted that they:

(a) only occurred where the regolith was thick enough for deep rotational slumps to occur entirely within the weathered mantle,

(b) were no longer active (the present dominant form of mass movement on the slopes he studied is debris slides, with no rotation),

(c) occurred on the upper parts of valley-side slopes.

Seepage heads were apparently closely associated with amphitheatres in the areas studied by Selby.
Selby (1967) provided the following description of seepage heads:

"...small hollows, usually with terracotted backwalls, from which originate swam plgy channels infilled with fine sediment. The origin of these features is not entirely clear but it is possible that the head is usually a small debris slide or shallow slump. The colluvial infill was derived from the head and is probably removed by slow seepage from the head which keeps the infill saturated. The seepage heads are only found on the midslopes and usually all occupy a similar position along any valley wall. This suggests that they result from seepage from a perched water table, which in a few cases may be controlled by a structural plane.... Many of the seepage heads occur within old debris slides as do many of the new debris slides." This description very closely fits the seepage heads of the South Taranaki terraces. Debris slides are also widespread, and closely associated with seepage heads (as described by Selby) on the terraces. A large debris slide is shown in Plate 6.11.

Amphitheatres in South Taranaki are not restricted valley sides, but also occur at the heads of many valleys on the terrace sequence (Plate 6.7). A spring at the base of a steep, terracotted headwall is characteristic of these valley-head forms; active slumping is occurring on the headwall, and valley-extension appears to occur via this process. The springs in these amphitheatres provide the major source of water in many terrace front valleys.

It is suggested here that valley extension by a process of headward sapping and slumping in amphitheatre valley heads is sufficient to explain the development of small terrace front valleys after cliff retreat ceases (Stage 3 above). Similar amphitheatre valley heads have been reported by Carter & Chorley (1960) in their study of
an expanding stream network on a river terrace in Connecticut USA; they were also envisaged by them as the growing tips of the channel network. Bunting (1961) has documented similar "headward sapping" by subsurface water in a U.K. study. As explained previously (p.151), amphitheatre slump heads and incipient valleys are not observed on the present coastal cliff because the rate of cliff retreat is much greater than their rate of headward retreat.

Valley-side and valley-head amphitheatres are less common east of Waverley on the terrace sequence, although the reason(s) for this are unclear at present. It is possible, however, that the greater thickness of non-marine cover beds (in particular of andesitic tephras) west of Waverley, may enhance conditions favourable for their formation.

It is suggested here that drainage network growth and bifurcation in South Taranaki largely occurs via headward retreat of favoured valley-side amphitheatres (insequent streams in classical terms eg. Cotton 1942, p.66). This would explain why many stream junctions are approximately at right angles, although Fleming (1953, p.68) pointed out that diversion by advancing coastal sand dunes could account for extensive shore-parallel drainage systems in parts of the terrace sequence.

In summary, valley-head and valley-side amphitheatres are therefore seen as the growing tips of stream networks on the South Taranaki terraces. The implication is that subsurface water is seen as a major factor in controlling the evolution of both slopes and stream networks in South Taranaki.
6.3 DATA COLLECTION

In the light of the foregoing discussion, the small terrace front valleys were selected for detailed study. Their small size meant that they could be adequately studied in the time available, their ages could be reasonably inferred to approximate the ages of the cliffs they dissect and they represent one of the simplest suites of fluvial landforms developed on the terraces. Fig. 6.3 shows a typical terrace front small valley; see also Plates 6.7, 6.8, 6.9.

Ideally, terrace front valleys could be studied on each terrace front (or cliff) of the terrace sequence. However, practically, this was not possible owing to:

(a) The time available; even five valleys from each of 10 terrace fronts would have been too large a task.
(b) Lack of suitable valleys on many terrace fronts eg. terrace width too narrow, vegetation too thick, base level control uncertain.
(c) Lack of stratigraphic information in the early stages of the study eg. the Rapanui, Inaha and younger unnamed terraces were not adequately mapped until near the end of fieldwork.

Data were collected from the Brunswick, Ngarino and Inaha cliffs (ie. the Kaiatea III Brunswick and Rapanui terrace fronts respectively) as follows:

(a) thalweg profiles surveyed with theodolite
(b) slope profiles surveyed with slope inclinometer (1.5m ground-surface length)
(c) terrace cover bed stratigraphy and wave cut surface position noted where possible in valley sides
(d) morphometric data measured from aerial photographs and survey data.

Twenty seven terrace front valleys were surveyed. Maps of all surveyed valleys are shown in Appendix 4, and their locations indicated on Map 1. Each terrace front valley is identified by a number, SV001 to SV027 inclusive. Not all the surveyed valleys are single, unbranched valleys eg. SV001 and 002 are part of a small
stream network (Plate 6.1), while other valleys such as SV015 have small bifurcations in the headwater tracts. Individual thalweg profiles are shown in Appendix 4, and are also plotted as a group in Fig. 6.4. The locations of all slope profiles is also given in Appendix 5. A total of 139 slope profiles were surveyed of which 54 were paired profiles on opposite sides of a single valley cross-section.

6.4 TERRACE FRONT VALLEYS

6.4.1 DESCRIPTION

The twenty seven terrace front valleys, which provide the primary data for this geomorphic study, are aligned approximately at right angles to the cliff fronts which they dissect (ie. shore-normal). They range in length from less than 100m to slightly more than 1200m, and are all inferred to have evolved largely through a headward sapping and slumping mechanism.

Typically, valley head amphitheatres are present at the upstream ends of the valleys and their tributaries (Plate 6.7). These are large basin shaped area with steep, terraceted headwalls, a basal spring(s) and swampy gently sloping floor, as described above. Occasionally the valley extends headwards of the major amphitheatre as a shallower depression (Plate 6.7), with no evidence of surface water except during, and immediately following, heavy storms.

Immediately downstream of the valley head, the valley floor is swampy, with no well defined stream channel present (Plates 6.9, 6.10). Material from the valley sides appears to infilled the valley floor. In the lower portions of most valleys (except the smallest),
their streams have incised a well defined channel which may be locally eroding directly into the underlying Tertiary mudrocks. Seaward of their exit through the cliff front, the valleys are incised below the present ground surface of the next youngest terrace. Larger valleys (eg. the confluence of SV001 and 002) are incised below the wave cut surface of this younger terrace; smaller valleys lie entirely within the cover beds. This next youngest terrace, particularly when broad, appears to act as a temporary base level for the dissecting streams (Plate 2.10). A well defined channel continues to be present on this younger terrace only for the larger streams; the smallest terrace front valley (SV021) drains directly onto the present ground surface of the younger terrace, and no channel or valley expression is evident.

Observations within valleys which retain a large proportion of their natural vegetation, (principally podocarp-hardwood-beech forest - see Fig. 2.8) eg. SV005, parts of SV002 and other valleys not included in the study, indicate that the surveyed valleys, which have been cleared for agriculture, are entirely similar to their forested counterparts except that:

(a) Mass movements (principally debris slides) are not as widely present in the forested catchments, which means that:
(b) less valley floor infilling is present, and
(c) streams have well defined channels for a greater proportion of their length.

Valley-side slopes are generally steep (c.30°), and rectilinear to slightly convex in profile form (Plate 6.8). Debris slides, seepage heads and amphitheatres are widely present within the terrace cover beds; spur and groove topography is well developed (Plates 6.6 and 6.9). Below wave cut surfaces amphitheatres are not present. Slopes steeper than c.20° typically display well defined terracettes (Plate 6.6); below this angle, terracettes are rarely present or poorly defined. Rotational slumps are occasionally present (Plate 6.12).
Stream flow in the valleys was continuous throughout the study period (January 1976 to January 1977), being largely derived from subsurface flow. The smallest valley (SV021) was the only valley with no stream flow evident; however, a small boggy area near the wave cut surface in its valley head remained persistently wet throughout the year. A small number of discharge measurements were made with a portable 90 degree V-notch weir (cf. Moore 1975) and indicated that stream discharge was approximately proportional to valley length. In SV001, measured discharge progressively increased downstream, indicating that the valley head springs are not the only source of stream flow in this valley. Visual estimates in other valleys support these observations. No quantitative estimate of the relative contributions of surface versus subsurface stream flow sources was attempted; however, my own qualitative assessment during the 12 months field season was that subsurface flow was the dominant source of stream in the survey terrace front valleys.

A limited number of specific conductivity measurements in SV001 showed that dissolved ion concentration changed little in a down valley direction (mean of 17 observations: 190 µmhos at 13°C). Dissolved load was therefore principally derived from subsurface sources. Conductivity was monitored continuously in SV001 for several weeks, and, as expected, showed a negative response to rainfall events.

6.4.2 GEOMETRY

Morphometric data are listed for 25 of the 27 surveyed terrace front valleys in Table 6.1. SV002 and SV003, which are magnitude 4 stream networks, are not included in the analysis because of their comparative complexity. Mean values are also stated for each measured variable in Table 6.1; geometric framework is shown in Fig. 6.3.
The following variables have been measured or calculated:

1. Valley age ($t$), which is approximated by the age of the cliff which the valley dissects.
2. Thalweg slice area ($V_{th}$), which is the difference between the present ground surface and the stream thalweg in the valley long-profile.
3. Mudstone thalweg slice area ($V_m$) i.e. the difference between the wave cut surface and the stream thalweg in the valley long-profile.
4. Mudstone ratio ($M$) i.e. $V_m/V_{th}$.
5. Valley area ($A$)
6. Valley length ($L_V$), which is the horizontal distance of the down valley theodolite traverse.
7. Valley fall ($F_v$), which is the vertical height difference between the valley head and the stream exit (strandline).
8. Stream length ($L_s$), i.e. the horizontal distance between the headwall spring and the stream exit (theodolite traverse).
9. Stream fall ($F_s$) i.e. the vertical height difference between the headwall spring and the stream exit.
10. Headwall height ($H$) which is the height difference between valley head and stream head or headwall spring i.e. ($F_v-F_s$)
11. Overland flow distance ($d_Q$) i.e. ($L_V-L_s$).
12. Shreve magnitude ($N$).
13. Relief ratio ($RR$) i.e. $F_v/L_V$.
14. Stream slope ($S$) i.e. $F_s/L_s$.
15. Mean valley width ($W$) i.e. $A/L_V$ (note that constant of channel maintenance is similar i.e. $A/L_s$).
16. Mean valley depth ($D$) i.e. $V_{th}/L_V$.
17. Drainage density ($DD$) i.e. $L_s/A$ i.e. reciprocal of constant of channel maintenance.
18. Mean cover bed thickness ($C$) i.e. ($V_{th}-V_m$)/$L_V$.
19. Cliff height ($CH$), which is the vertical height difference of the wave cut surface above and below the cliff (terrace front).
20. Mean stream incision or downcutting rate ($r_d$) i.e. $D/t$.
21. Mean headward retreat rate ($r_h$) i.e. $L_V/t$.
22. Normalised thalweg slice ($V_{th}/L_s^2$).
23. Uplift rate of strandline at valley exit ($U$).
A correlation matrix for the morphometric variables is given in Table 6.2, and indicates that many variables are highly correlated. Power curve fits were investigated, and resulted in correlations that were similar to the linear regressions in most cases. The geometry of the surveyed valleys is therefore seen as similar, in broad terms, for each valley.

No attempt will be made here to analyse in detail all the interrelationships evident in Table 6.2. Since I am principally concerned with changes in time (or evolutionary aspects) of the terrace front valleys, I will focus attention on the variables.

(1) valley age (t)
(2) area of thalweg slice ($V_{th}$)
(3) rate of stream downcutting ($r_d$)
(4) rate of headward retreat ($r_h$)

Somewhat surprisingly, valley age (t) is strongly correlated with few other variables. Valley age is most strongly correlated with mean cover bed thickness and valley fall ($p(H_o)<0.001$) as well as with stream fall, mean valley depth and drainage density ($p(H_o)<0.01$). The significant correlation between valley age and mean cover bed thickness is expected in view of the approximately linear increase of non-marine cover bed thicknesses indicated previously (Fig. 6.1C); while the correlation with depth of incision measures (stream fall, valley fall and mean valley depth) suggests the unsurprising trend of progressive valley deepening with time. On the other hand, the significant negative relationship with drainage density suggests a trend towards decreasing drainage density with time. (cf. Fig. 6.1B which indicates the opposite trend on a terrace-wide basis). Note, however, that within-valley drainage density (DD) does not necessarily relate in a simple manner to stream density measured on a terrace wide basis.
As indicated by Chappell (1974a), a simple measure of stream incision is the difference between the initial groundsurface and the present position of the stream thalweg - this he referred to as the area of thalweg slice. In this study, the area of thalweg slice is taken as the difference between the present adjacent interfluve groundsurface and the present thalweg. The reasons for this definition is that since deposits such as tephras and dunesands (which comprise a major proportion of the non-marine cover beds in the valleys studied here), could reasonably be assumed to mantle a landscape uniformly, such accreting deposits in interfluve areas must also have been eroded by stream activity in the valleys. Consequently, while the initial groundsurface can be approximated to the top of the marine cover beds, the actual volume of material eroded by the stream is more closely approximated relative to the present groundsurface. The area of thalweg slice is therefore taken to represent the sum total of thalweg lowering along the valley profile since valley initiation.

Intuitively one might expect the following factors to control thalweg lowering on the South Taranaki terraces, regardless of the processes acting:

1. The time elapsed since valley initiation, because stream incision is likely to increase progressively with time (see valley age discussion above).
   Variable: valley age (t)

2. Rock type variations.
   Variables: average cover bed thickness (C), proportion of mudrocks in thalweg slice area (M).

3. Available relief and initial slope.
   Variables: average cover bed thickness (C), cliff height (CH), valley fall (Fv) and relief ratio (RR).

4. Stream discharge and headwall spring discharge.
   Variables: valley length (Lv) and valley area (A), since higher discharges are associated with larger valleys.

   Inspection of Table 6.2 confirms that C, Fv, RR, Lv and A are all significantly correlated with thalweg slice area (p(H0)<0.01)
while the correlations with $t$, $M$ and $CH$ are not significant ($p(H_0)<0.01$). The dependence of $V_{th}$ on these variables was investigated using linear multiple regression techniques (Draper & Smith 1966; Neter & Wassermann 1974). Best fits were obtained using log-log multiple regressions. A simple correlation matrix is given in Table 6.3 for the eight log-transformed variables listed above. Sample regression equations are also listed in Table 6.3.

A best fit log-log multiple regression containing all eight variables explains 99.4% of the variance in thalweg slice area (Table 6.3). Unfortunately the intercorrelation (multicollinearity) amongst the "independent" variables (see Table 6.2), makes interpretation of the implied relationships, in this and the other best fit regressions listed in Table 6.3, somewhat difficult. [Note here that while the effects of intercorrelation do not prevent a good fit being obtained, true cause-effect relationships are certainly masked (Neter & Wassermann 1974, p.341)]. In addition some of the variables make an insignificant contribution to the overall explanation of variance in $V_{th}$ (see Table 6.3). eg. a best fit regression containing only 3 of the 8 variables (valley area, valley fall and relief ratio) still explains 99% of variance in $V_{th}$, while a best fit regression of two variables (valley fall and relief ratio) explains 98% of variance.

The problem of multicollinearity is not easily circumvented. As Neter & Wassermann (1974, p.346) point out, even the dropping of intercorrelated variables from the regression does not necessarily remove their effect if they are still correlated with the remaining variables. ie, although the two variable regression above contains valley fall and relief ratio, which are not significantly correlated with each other ($p(H_0)>0.05$), they are still affected by strong correlations with excluded variables eg. between valley fall and valley length ($p(H_0)<0.001$).
The implication of the above discussion is that a further search for independent variables, and/or a closer scrutiny of those selected, is required. For example, valley area is strongly correlated with valley length simply for geometric reasons so that both should not be included in the same regression. Similarly mean cover bed thickness and valley fall, which are strongly correlated with valley age, are not truly independent variables. In this way many potential regressions may be eliminated from consideration. The result is that a simple two variable regression containing valley length and valley age (no significant intercorrelation; \( p(H_0) > 0.1 \)), which explains 95.6\% of variance in \( V_{th} \) (Table 6.3) or a three variable regression containing valley length, valley age and mudstone ratio (96.8\% of variance explained and no significant \( p(H_0) > 0.5 \) intercorrelation between included variables), are seen as most satisfactory.

The presence of valley length (which by itself can account for 93\% of variance in \( V_{th} \)) in both the above regression equations might be explainable simply for geometric reasons ie. larger valleys must have larger thalweg slice areas, other factors being equal. To eliminate the effect of this possible relationship, multiple regression analysis using normalised thalweg area \( (V_{th}/L_V^2) \), instead of \( V_{th} \), was investigated (Table 6.4). In this case the contributions of valley age, valley length and mudstone ratio are still significant \( (p(H_0) < 0.001) \), although only 86\% of variance in \( V_{th}/L_V^2 \) is explained by these three variables. Note that the strong correlation between \( V_{th}/L^2 \) and relief ratio (see Table 6.4) is expected since \( V_{TH}/L_V = \) average depth \( (D) \) and \( D/L \) (ie. \( V_{TH}/L_V^2 \)) is therefore a gradient term.
Another simple measure of vertical incision is mean valley depth ($D$), which when divided by valley age, yields an average rate of stream incision ($r_d$) for the catchment. Inspection of Table 6.2 shows that $r_d$ is significantly correlated with $V_{th}$, $V_m$, $A$, $L_v$, $L_s$, $F_v$, $F_s$, $W$, $D$, $D_D$ and $r_h$ ($p(H_0)<0.001$) and with $V_{th}/L^2$, $M$, $RR$ and $S$ ($p(H_0)<0.01$). A similar measure of valley development is valley length divided by valley age which represents a mean rate of headward retreat ($r_h$). These and other measures of erosion rate are discussed more fully in Chapter 7.

I conclude this discussion of valley geometry with the unsurprising statement, that valley development is controlled by a complicated interrelation of many factors. However, two of these factors - valley length (a surrogate for stream discharge) and valley age - are identified as being of major importance, and together explain almost 95% of variance in thalweg slice area ($V_{th}$). Although valley length is expected to increase progressively in time (cf. headward retreat arguments earlier in this chapter), the surveyed valleys were deliberately selected to suppress this relationship; valley length and valley age are therefore considered to be independent of each other within the suite of terrace front valleys considered here.

6.5 STREAM THALWEGS

Surveyed long profiles of stream thalwegs are shown for each small valley in Appendix 4; complete numeric data are also given. Figure 6.4 shows the small valley long profiles plotted as a group. Most profiles are gently concave upwards with the notable exception of SV002A and SV003, and to a lesser extent SV001. Standard curve fitting techniques showed that individual long profiles could be adequately fitted by linear, power, exponential and logarithmic
functions; however, a power curve of the form \( Y = AX^B \)
(where \( X \) = vertical fall from valley head and \( Y \) = horizontal distance,
in metres) most consistently gave excellent fits. Best fit power
functions are given in Table 6.5 for each stream thalweg; sample fits
are shown in Fig. 6.5.

Inspection of the valley long profiles (Appendix 4, Fig. 6.4)
and the power function curve fits (Table 6.5, Fig. 6.5) indicates that
for individual small valleys:

(1) There is a noticeable lack of expression of the wave cut
surface in the long profile, in most instances. For changes
in lithology, such as occurs across the wave cut surface, the
presence of knickpoints might be expected.

(2) While vertical fall increases with distance from the valley
head, the depth of incision below the present ground surface
does not necessarily increase in the same manner. In
particular the maximum depth of incision does not always
coincide with the downstream end (cliff line) of the valley.
Furthermore, the depth of incision decreases markedly once
the stream passes the cliff line and flows across the next
younger terrace. (This younger terrace appears to act as a
temporary base level, particular when it is wide: surveyed
valleys were selected wherever possible so as to utilise this
relationship i.e. the evolution of surveyed terrace front
valleys was largely independent of base level changes.)

Although individual thalweg profiles are very closely described
by individual power curves, the correlation matrix (Table 6.2) and the
multiple regression analysis (Tables 6.3 and 6.4) presented in the
previous section, clearly indicate that factors other than horizontal
distance are influencing vertical incision of the valleys as a group.
This observation is further reinforced when a single power curve is
fitted to all the survey data, and gives a comparatively low correlation
\( R^2 = 0.72 \) for the relationship between stream fall and distance
downstream.

Many other workers have discussed the mathematical representa-
tion of stream long profiles e.g. Shulits 1941, Yatsu 1956, Hack 1957,
Broscoe 1959, Tanner 1971, Connelly 1972. It seems from their work that
no single mathematical function adequately describes all stream
profiles, and the reasons for this obviously lie within the complex process-form relationships within the fluvial system. The small valley suite studied here confirms this observation and suggests that even for a small number of very similar streams within a relatively uniform environment (climatic, geological), many of the same complexities prevail. However, the advantage of the South Taranaki situation lies with the fact that many of these "complexities" can be identified more closely than in previous studies.

On the other hand, the concave-upward form of stream long profiles appears sufficiently widespread for many workers to consider it as the "equilibrium form" towards which most rivers tend (eg. Mackin 1948, Langbein & Leopold 1964, Connelly 1972). Connelly (1972), following Langbein & Leopold (1964) has argued that a long profile of the form

\[ Y + \text{constant} = -K \ln X, \]

where \( Y = \text{altitude} \)
\( X = \text{horizontal distance from source} \)

corresponded to a profile of minimum work, and that this represented the equilibrium profile towards which all streams should tend. I note here that Hack (1957) derived the same equation using the empirical relationship

\[ S = 25M^{0.6}/X \]

where \( S = \text{slope} \)
\( M = \text{median diameter of bed materials} \)
\( X = \text{distance downstream} \)

which for the case where particle size remains constant with distance
(a reasonable assumption perhaps for short stream), reduces to the logarithmic relationship above. Further comment will be made on this subject in the following chapter.

6.6 VALLEY SIDE SLOPES

One hundred and twenty eight individual slope profiles were surveyed using a slope inclinometer with a groundsurface unit length of 1.5m (cf. Pitty 1967), which could be read to the nearest 0.5°. Of these fifty four profiles were paired profiles ie. on opposite sides of the valley from a single point on the stream thalweg.

Slope profiles were sited randomly within the small valleys as far as possible (cf. Pitty 1966) although vegetation patterns, human disturbance (eg. farm tracks) and surface irregularities imposed constraints in some instances. Profiles were measured from stream channels, upslope and for some distance onto the undissected terrace surface. As far as possible, profiles were measured orthogonal to the slope contours. A constant groundsurface length of 1.5m was used for all measurements and was not varied. No attempt was made to break profiles into segments except to define where the wave cut surface intersected the profile.

For the purposes of analysis, some difficulty was encountered in defining the upper and lower boundaries of the slope profiles. For example, while it could be argued that the stream channel itself forms the logical lower boundary, account must also be taken of variations in the width of the valley floor. Furthermore, as indicated previously (see Plate 6.10 for example), the stream channel is poorly defined on some instances, so that unambiguous placement of the lower boundary would not always be possible. For the purposes of analysis here, the
lower boundary of the slopes was taken at the point of maximum curvature in the lowest portion of the slope profile; this point presumably represents a point which roughly separates dominantly slope processes from dominantly valley floor processes. Similarly, a well defined upper inflection point was identified as separating dominantly slope processes from dominantly interfluve processes. Defined in this way, the number of individual slope angle readings is approximately 5000 in this study. A simple histogram plot of all slope data is shown in Fig. 6.6. Mean slope angle is 28.2°. Some typical slope profiles are also shown in Fig. 6.6.

Slope data was analysed in the following groups in turn:

(1) Total population.
(2) According to valley age.
(3) West versus east facing slopes.
(4) Slope segments developed in cover beds versus slope segments developed in mudstone (below wave cut surface).

A summary of this analysis is given in Table 6.6 from which the following conclusions may be drawn:

(a) East facing slopes are on the average 1.1° steeper than west facing slopes \( (p(H_0)<0.001) \).

(b) Cover bed slope segments are on the average 4.6° shallower than slope segments developed in mudstone \( (p(H_0)<0.001) \).

(c) On average, approximately 60% of the slope (measured in a vertical direction) lies within the cover beds, and 40% below the wave cut surface (mudstone).

(d) The mean slope angle is not significantly different between the 200 and 300Ka age group valleys \( (p(H_0)>0.05) \).

(e) On the average 300Ka age group slopes are 3.5m higher and 7.4m longer (horizontal distance) than the 200Ka age group \( (p(H_0)<0.001) \) ie. 8.5m longer groundsurface length (see Table 6.6):.

(f) All slope distributions are positively skewed.

In addition, variation in mean slope angle with position in the catchment (expressed as distance downstream from valley head) and with profile height were investigated (Fig. 6.7). The relationship
between mean slope angle and distance downstream is poorly defined: a best fit power curve has a correlation coefficient \( R^2 = 0.24 \) although the correlation is significant \( (p(H_0)<0.001) \). The relationship between mean slope angle and profile height (valley depth) is similarly poorly defined \( (R^2 = 0.19) \).

6.6.1 SLOPE ASYMMETRY

Strictly speaking, an asymmetrical valley is one in which the opposing sides are not exactly mirror images of each other about the valley axis. However, as Kennedy (1976) has pointed out, when geomorphologists speak of asymmetry they usually refer to substantial differences in overall slope geometry or in slope steepness on opposite valley sides. Unfortunately there is no universally accepted definition of "substantial" differences. Some workers (eg. Hadley 1961, French 1971) have advocated the use of asymmetry indices based on slope (mean or maximum) angle ratios between opposite valley sides. However, the arbitrary nature of the selection of the critical ratio, and the obvious variation in its meaning with changing mean profile steepness, makes its use of somewhat doubtful validity.

In the previous section I showed that on the average east facing slopes are 1.1° steeper than west facing slopes in the small terrace front valleys and that this was a statistically significant \( (p(H_0)<0.001) \) difference. In this section I now focus attention on individual paired slope profiles, and test the hypothesis that either the east or west facing slopes are steeper, for each of 50 pairs, using the Mann-Whitney U-test (see Siegel (1956) for discussion). A significance level of 95% was adopted for rejection of the null hypothesis (ie. \( p(H_0)<0.05 \)). Results are summarised in Table 6.7 and indicate that in 31 cases there was no significant difference, in 5 cases the
west facing slope was significantly steeper and in 14 cases the east facing slope was significantly steeper.

Kennedy (1976) has listed the following possible causes of valley asymmetry.

(1) Coriolis force
(2) differences in microclimate
(3) differences in slope dimensions
(4) variable lithology
(5) geological structure
(6) warping
(7) evolution of the drainage net
(8) glaciation.

The operation of the Coriolis force, which has the effect of deflecting moving objects to the left in the Southern Hemisphere, was postulated as a possible cause of river meanders and valley asymmetry by many early workers eg. Gilbert 1884, Hilgendorf 1906, Davis 1908, Einstein 1926, Exner 1927. However, more recent studies have tended to dismiss it as a significant factor eg. Fairchild 1932, Currey 1964. If the Coriolis force was a significant cause of valley asymmetry in South Taranaki, deflection of moving bodies of water to the left should result in steeper left (west facing) valley side slopes. That the opposite is observed therefore suggests that the operation of the Coriolis force is not a significant factor in the development of valley side slopes in South Taranaki.

Differences in microclimate, particularly in response to variation in solar radiation, moisture and wind with changing aspect have been suggested by many workers as a likely cause of valley asymmetry eg. Hack & Goodlet 1960, Currey 1964, Bik 1968, French 1971. In particular,
in northern hemisphere studies, north facing slopes have often been shown to be significantly steeper than south facing slopes (summarised by Kennedy 1976), apparently in response to microclimate variations. In the South Taranaki case, where the valley side slopes are oriented east-west, the situation (particularly with regard to radiation input) is perhaps not as clear cut. Certainly the predominance of westerly winds (see 2.7 Climate) and their resultant influence of rainfall variability between E-W oriented slopes might be a possible source of soil moisture variations significant enough to produce the observed asymmetry; the microclimatic factors cannot therefore be overlooked.

Differences in slope dimensions, evolution of the drainage net, variability in lithology and glaciation appear to be ruled out of contention in South Taranaki; however, the effect of geological structure has already been indicated as a major cause of asymmetry in shore-parallel valleys (ie. N-S oriented valley side slopes) on the South Taranaki terraces (Plates 6.3 and 6.4). Here the effect of increased subsurface discharge on the updip (south facing) valley side has resulted, in general, in lower mean slope angles than on downdip (north facing) valley side slopes. While Kennedy (1976, p.191) assumed that "any manner of structural influence upon the cross-profiles of valleys cut in essentially homogeneous rocks will cease to operate at dip angles of 1° or less), the wave cut surface dips are all less than 1° for the terrace front valleys studied here and indicate that her assumption is not of universal validity. Since the valley-side slopes studied in South Taranaki terrace front valleys are oriented roughly normal to the wave cut surface dip, the effect of structure is thought to be eliminated.

The effects of tilting or warping of the landsurface in relation to valley asymmetry have received little attention in the literature. However, the distribution of isostatic deformation in
northern hemisphere continental areas for example, seems sufficiently widespread for this factor to be considered more seriously. Nanson (1980) has reported a case from western Canada in which an easterly bias to channel migration is seen as the response to post-glacial isostatic tilting of 0.0003 to 0.0004 in the same direction (ie. of c.0.002°). In a previous chapter I have reported shore-parallel deformation on the South Taranaki terraces amounting to 0.08° on the Ngarino strandline and 0.10° on the Brunswick strandline between Hawera and Wanganui; direction of tilt is to the west. Since all the surveyed valleys occur in this region of the terraces (see Map 1 for location), it is a possible source of valley asymmetry; a westerly tilt should result in steeper east facing slopes, which is observed.

I conclude that either microclimate and/or tilting-type effects are the most likely causes of the observed valley asymmetry in South Taranaki.

6.7 CONCLUDING REMARKS

The small terrace front valleys have been quantitatively described, and simple notions regarding their evolution have been introduced. In particular, it is clear that the geological information presented in the preceding chapters is critical to an understanding of the factors important in valley extension and deepening, and valley side slope evolution.

I concluded in Chapter 1 that one of the major shortcomings of quantitative evolutionary models in geomorphology is the lack of adequate testing against natural landscapes. However, even the most comprehensive recent models are only beginning to grapple with a three dimensional landscape (eg. Smith & Bretherton 1972, Ahnert 1976,
Armstrong 1976). Most current models concentrate on portions of a landscape and depict their evolution in profile form (ie. two dimensions) eg. the quantitative slope development models listed in Chapter 1 (p.3). The stream thalweg data and valley side slope data discussed in this chapter relate to this latter category of models, and enable a range of two dimensional evolutionary models to be tested in the following chapter.

Although not specifically modelled, the three dimensional evolution of the terrace front valleys may be visualised since the thalweg and slope profiles are roughly orthogonal. Furthermore the volume of material removed from a valley since its initiation is readily apparent in relation to the terrace groundsurface. Thus I introduce the quantitative evolutionary models in the following chapter via a discussion of erosion rates.
7.1 EROSION RATES

Quantitative measurements of erosion rates are normally derived in one of four major ways:

1. Measurement of present day stream loads eg. Corbel (1959); Douglas (1969)
2. Estimation of the amount of sediment deposited in dams or closed basins eg. Gordon (1979)
3. Direct measurement of present day ground lowering using "erosion pins" or similar device eg. High & Hanna (1970)

All of the above methods are subject to measurement and interpretation errors, although the method of reconstruction of initial forms appears best suited to long term estimates (see Chapter 1). Long term estimates of erosion rate based on sediment accumulation in closed basins suffer from two major drawbacks:

a. Assessment of trap efficiency (Brune 1953).

b. The exact source of the sediment is unknown, and is usually assumed to be uniformly removed from the catchment area (eg. mean ground lowering in mm/Ka) - see Trimble (1977) for discussion. Mead (1969) reviewed the errors involved in calculating denudation rates based on modern stream load data. Since many of these relate to human interference, similar errors presumably relate to methods involving direct measurement of present ground lowering.

Two measures of erosion rate, based on the reconstruction of initial forms, were introduced in Chapter 6: these were mean rate of stream downcutting ($r_{d}$), and mean rate of headward retreat ($r_{h}$) for individual small valleys, both expressed in cm/Ka (Table 6.1). The overall mean rate of headward retreat is $c.220$cm/Ka, while the overall mean rate of downcutting is $c.12$cm/Ka. Thus if valley extension has occurred progressively headwards from the terrace front cliffs as described in Chapter 6, then the rate of headward retreat
has far exceeded the rate of downcutting. Note also that the present mean rate of cliff retreat \( (r_c) \) on the modern coastline is c.70cm/year or 70m/Ka, which is more than an order of magnitude greater than the implied mean rate of headward valley retreat \( (r_h) \) for the surveyed terrace front valleys and modern coastal cliff*,

\[ r_c \gg r_h \gg r_d \]

of the two erosion rate measures \( (r_d \text{ and } r_h) \) for the terrace front valleys, the derivation of \( r_d \) is probably more useful: while both express mean rates of valley development, variation in stream downcutting within a catchment is expressed at any point by the vertical height difference between the stream thalweg and the undissected adjacent interfluves.

Consider four simple models of headward retreat as follows:

(A) If no significant headward retreat had occurred in the valleys, and downcutting was therefore the principal cause of valley enlargement, then mean rates of downcutting could be calculated for all points along the present thalweg based on the height difference between the terrace surface and stream thalweg. In this case valley age remains constant for all points along the thalweg. The gentle concave upward long profiles of the valleys therefore suggest that downcutting rates decrease upstream (Fig. 7.1A).

(B) Assuming a constant rate of headward retreat, downcutting rates can be calculated as above by linear interpolation of thalweg age between valley head (zero age) and the cliff front (valley initiation age eg 200Ka BP). This model implies that downcutting rates increase upstream for concave upwards thalweg profiles (Fig. 7.1B).

(C) If the rate of headward retreat increases with time, downcutting rates increase upstream at a greater rate than model B for concave upwards thalweg profiles (Fig. 7.1C).

(D) If the rate of headward retreat decreases with time, depending on the exact pattern of this decrease, the rate of downcutting would decrease upstream (rapid deceleration), or if the rate of headward retreat had a similar form to the depth of incision curve D(L) - see Fig 7.1 - the rate of stream downcutting would remain relatively constant upstream (Fig. 7.1D).

* As indicated in Chapter 6, the small terrace front drainage systems are initiated when cliff recession slows, and finally ceases at the culmination of transgression cycles ie. when \( r_h > r_c \). This accounts for their absence on the modern coastal cliffs.
Of the above models, model A is not consistent with the field description and interpretation of headward retreat (Chapter 6), models B and C seem unlikely for theoretical reasons (eg. \( r_d \propto L_v \) - see Table 6.2 - ie. \( r_d \) related to discharge), while model D is consistent with field evidence and seems theoretically reasonable. However, it should be noted that since the thalweg profiles of the small terrace front valleys display only a gentle upward concavity, in most instances, Model B is a close approximation to Model D.

Erosion rates are generally reported in the literature in a variety of units and dimensions (cf. Caine 1976). However, two of the most common are:

1. The mass of volume of sediment transported out of a river catchment (eg. expressed as kilograms or cubic metres/unit time).
2. Areal averaging to give a mean rate of lowering of the landsurface (eg. expressed in mm/unit time).

For the South Taranaki terrace front valleys, the volume of material removed is simply estimated as

\[
\text{volume (v)} = \left( \frac{V_{th}}{2} \cdot L_v \right) \cdot A
\]

where

- \( V_{th} \) = area thalweg slice
- \( L_v \) = valley length
- \( A \) = valley area,

where the valley side slopes are taken to be rectilinear from stream thalweg to terrace surface (ie. V-shaped cross-section). The erosion rate is therefore \( v/r \) expressed as cubic metres/Ka. To convert this to an areally averaged mean rate of groundsurface lowering (\( r_g \)),

\[
\begin{align*}
\text{\( r_g \)} & = \frac{v}{A} \\
& = \frac{V_{th}}{2}L_v \\
& = \frac{r_d}{2}
\end{align*}
\]

In other words the mean rate of groundsurface lowering is approximately half the mean rate of stream downcutting ie. \( r_g = 6\text{cm/Ka} \), given the assumption of essentially rectilinear valley side slopes.
To investigate this areally averaged mean rate of ground-surface lowering, consider the following four simple models of valley widening (Fig. 7.2), which are similar to the four models of headward retreat considered above: (for convenience the slopes are assumed to be essentially rectilinear).

(A) If the slope divide is assumed to be fixed i.e. no valley widening is occurring, the mean rate of slope lowering must decrease upslope, since slope age remains constant across the slope profile (Fig. 7.2A).

(B) Assuming a constant rate of valley widening, the mean rate of slope lowering \( r_g \) must remain constant across the slope (Fig. 7.2B).

(C) If the rate of valley widening is accelerating with time, the rate of slope lowering must increase upslope (Fig. 7.2C).

(D) For a decelerating rate of valley widening, the rate of slope lowering must decrease upslope (Fig. 7.2D).

Of the above models (all of which assume essentially rectilinear valley side slopes (see Fig. 6.6B), model A seems inappropriate because of the demonstrated strong positive correlation between valley depth and valley width (Table 6.2). Model B is plausible, although only if mean slope angle remains relatively constant with increasing depth, which in turn indicates a constant rate of stream downcutting (see discussion above of headward retreat models). Model C seems theoretically unlikely, while model D is possible for the case of constant stream downcutting and progressive increase in slope angles (cf. Fig. 6.7).

In view of the previous discussion regarding models of valley headward retreat, the simplest model appears to be one of

(a) constant stream downcutting (ie. decelerating headward retreat)

(b) decelerating valley widening

(c) decreasing rates of slope lowering upslope.
On the other hand, the gentle upward concavity of stream thalwegs relative to their terrace surfaces indicates that a constant headward retreat model may be a close approximation to the constant stream downcutting model (D) above. In the case of valley widening, if mean slope angles remain constant (constant slope geometry, regardless of rectilinear, or otherwise, slope profiles), a constant widening model (B) may be appropriate for valley side slopes. Consequently for this simplified model of
(a) constant headward retreat rates
(b) constant stream downcutting rates
(c) constant valley widening rates
(d) constant slope lowering rates,
all points within individual terrace front valleys would be lowering at approximately the same rate ie. the terrace front valleys would be in a situation of dynamic equilibrium or steady state (cf. Hack 1960).

The foregoing discussion provides a general framework for the development of some quantitative models of valley evolution in South Taranaki. However, before considering the models in detail, some mention must be made of the mathematical representation of geomorphic processes within the South Taranaki terrace front valleys. This topic is introduced by a general summary of previous relevant overseas studies in the following section.

7.2 MATHEMATICAL REPRESENTATION OF GEOMORPHIC PROCESSES IN DETERMINISTIC MODELS

7.2.1 PREVIOUS WORK

Young (1963) observed that for valley side slopes, two major classes of processes could be recognised:
(1) Those involving instantaneous removal of material
(2) Those involving slow point-to-point downslope transfer.
For processes involving point-to-point downslope transfer, Young argued that the rate of transfer could be:

(a) proportional to the sine of the slope angle, since the downslope component of gravitational force acts parallel to the ground surface in this way e.g. soil creep, or

(b) proportional not only to \( \sin \alpha \), but also distance downslope \((x)\) e.g. if the volume of the transporting agent increases downslope, as for slope wash or overland flow.

In the first case (soil creep), ground surface lowering is described by the following equation:

\[
\frac{dz}{dt} = -K_1 \frac{d^2z}{dx^2}
\]  

which is the well known diffusion equation in 1 dimension.

(geometric frame of reference shown in Fig. 7.3, where \( z \) is the vertical axis and \( x \) the horizontal axis; \( K_1 \) is a constant).

In the second case above (overland flow) ground surface lowering is given by

\[
\frac{dz}{dt} = -K_2 \frac{d^2z}{dx^2} = -\frac{d}{dx} (K_2 x \frac{dz}{dx})
\]  

For processes of instantaneous removal, Young argued that ground lowering might be

(a) constant, regardless of slope angle

(b) proportional to sine of the slope angle

ie.

\[
\frac{dz}{dt} = K_3 \quad (3a) \quad \frac{dz}{dt} = -K_4 \sin \alpha \quad (3b)
\]

(making the small angle approximation \( \sin \alpha \approx \tan \alpha \)).

"Processes of direct removal" were thought to describe ground surface lowering by solution, by Young (1963), although he gave no evidence to support such an assumption. In particular, I note that the gravitational force is less relevant to solution processes than other factors such as residence time and temperature and composition.
of the transporting medium, which in turn control its aggressiveness.

At about the same time as Young was developing the above models, Scheidegger (1961, 1964) developed models along similar lines. Scheidegger (1961) presented the following three models of ground-surface lowering:

\[ \frac{dz}{dt} = -K_5 \]  
\[ \frac{dz'}{dt} = -z \]  
\[ \frac{dz}{dt} = -\frac{dz}{dx} \]  

but did not equate them with specific geomorphic processes. (Note that equation (6) is similar to equation (3b)). However, as pointed out by Scheidegger himself, these models, while appealing for their simplicity, do not accurately represent the course of slope development, because processes supposedly act normal to the slope. Thus vertical lowering is represented by "the vertical effect of the weathering action" (Scheidegger 1970, p.136). In this way (see Fig. 7.3), equations (4), (5) and (6) above become

\[ \frac{\partial z}{\partial t} = -K_5 \sqrt{1 + \left( \frac{\partial z}{\partial x} \right)^2} \]  
\[ \frac{\partial z}{\partial t} = -z \sqrt{1 + \left( \frac{\partial z}{\partial x} \right)^2} \]  
\[ \frac{\partial z}{\partial t} = -\frac{\partial z}{\partial x} \sqrt{1 + \left( \frac{\partial z}{\partial x} \right)^2} \]  

all of which are non-linear partial differential equations.

Equation (9) was developed by Scheidegger (1964) to allow for variation in rock type in a simple rectilinear initial slope, as follows:

\[ \frac{\partial z}{\partial t} = -\alpha(x,z) \frac{\partial z}{\partial x} \sqrt{1 + \left( \frac{\partial z}{\partial x} \right)^2} \]  

where \( \alpha(x,z) \) is a lithological resistance factor.
I note here that equations (9) and (10) were not envisaged by Scheidegger as describing any specific process, but rather as describing the overall effect of many contributing processes.

Nearly all more recent deterministic models of slope development follow the general principles outlined in the models of Scheidegger and Young eg. Hirano 1968, 1975; Kirkby 1971, 1976; Carson & Kirkby 1972; Chappell 1974a; Mizutani 1974; Ahnert 1973, 1976; Armstrong 1976.

The detailed model of Ahnert (1976) is a good example of the general approach; he incorporates the following individual processes: splash erosion, overland flow, plastic flow, viscous flow and debris slides, each of which has a separate mathematical formulation in his model. The quantitative formulation of each process is derived partly from empirical and partly from theoretical studies as follows:

1. splash erosion, rate of transport, $\sin \alpha^n$ (eg Mosley 1973, Moeyersons & De Ploey 1976),
2. overland flow, rate $\alpha x \sin^m$ plus erodibility factor (eg. Zingg 1940, Musgrave 1947, Horton 1945),
3. plastic flow, rate $\alpha \sin \alpha$, cohesion and soil thickness (eg. Souchez 1966),
4. viscous flow rate $\alpha \sin \alpha$, fluidity coefficient and soil thickness (eg. Kirkby 1967, 1971),
5. debris slides, crudely approximated by instantaneous downslope distribution when slope angle exceeds some critical value (45° in Ahnert's study) cf. in the New Zealand context James (1973).

In addition to the above processes, Ahnert's model incorporated the effects of weathering and baselevel changes. In particular this latter factor is of great importance to slope modelling and three categories are generally recognised

1. basal accumulation eg. screes, alluvial fans
2. complete removal of erosion products
(3) removal and incision at
   (a) constant rate
   (b) accelerating rate
   (c) decelerating rate

7.2.2 A DETERMINISTIC MODEL OF TERRACE FRONT VALLEY DEVELOPMENT IN SOUTH TARANAKI

Using the data described and analysed in previous sections, two major components of valley evolution are investigated here:
   (a) stream thalweg development
   (b) valley side slope development.

For a complete three dimensional picture of terrace front valley evolution, the areal development of the valleys (ie. stream network development in plan view) should also be studied. However, this aspect of valley development has yet to be approached deterministically, and appears not to be amenable to the kind of analysis presented here.

In view of the conclusion of Moss & Walker (1978), that slope processes are best viewed as extensions of fluvial processes, it seems appropriate that a single, general model be applied to both. The model described here is an extension of that described by Chappell (1974a).

Following Scheidegger (1961), groundsurface is represented by

$$\frac{\partial z}{\partial t} = \phi(x,z) \sqrt{1 + \left(\frac{\partial z}{\partial x}\right)^2}$$

(11)

where $\phi(x,z)$ represents the action(s) of any given process(es) which are assumed to act normal to the groundsurface (see Fig. 7.3). Three processes are included in the model: solution, corrasion and mass-movement. An option for headward retreat and basal lowering are also included, as well as provision for simple variation in rock type.
(1) **Solution processes**

Chappell (1974a) used an equation of the form

$$ \phi(\text{solution}) = A(x+d)e^{-\frac{T}{x}} $$

(12) where

- \( A = \) constant
- \( x = \) stream length
- \( d = \) overland flow distance
- \( T = \) solution time constant

This equation was used to model the solution process within valleys on the coral terraces of Huon Peninsula (bedrock almost pure \( \text{CaCO}_3 \)). In the case of the South Taranaki terraces, the solution process differs from the Huon Peninsula in that:

(a) The bedrock is not \( \text{CaCO}_3 \), and many more ionic species are likely to be important in dissolved loads of streams.

(b) Much of the stream discharge originates in valley head and valley side springs, whereas on the Huon terraces subsurface flow is apparently a negligible source of stream flow in the valleys studied by Chappell (1974a).

The limited number of conductivity measurements which I performed in SV001 and SV004 indicate that for base flow, little change in dissolved ion concentration occurs in a downstream direction. This suggests that for base flow, the bulk of dissolved load originates from subsurface sources. Thus only during storm events when a substantial contribution from surface flow occurs, is significant solution of the stream bed likely to occur. In this situation, equation (12) above is probably a crude approximation i.e. aggressiveness tends to increase downstream as water volume increases, but tends also to simultaneously decrease as travel distance increases.

A similar situation is thought to apply for valley side slopes, where once again subsurface flow appears to be the predominant source of overland flow cf. wave cut surface springs.

(2) **Corrasion processes**

Empirical relationships established by Zingg (1940), Horton (1945), Musgrave (1947), Emmett (1970) and others suggest that the corrasion process of groundsurface lowering is related to discharge,
slope angle and some erosivity function. For a uniform rock type, and assuming that discharge increases linearly downslope, a first approximation is given by

$$\phi_{\text{corrision}} = B \times \sin \alpha$$  \hspace{1cm} (12) \hspace{1cm} \text{where}

- $B =$ constant
- $x =$ horizontal distance
- $\alpha =$ slope angle

This model roughly accords with my observation that stream channels in the terrace front valleys are incised in slot-like paths in their downstream portions, while for valley side slopes, spur and groove topography is best developed in the lower portions of the slopes.

(3) Mass movement processes

These are clearly important in the evolution of the small terrace front valleys: slumping in valley heads and on valley side slopes, and the presence of terracettes and other mass movement scars (eg debris slides) are ample evidence of this. In addition, the swampy channel infill in the headwater tracts of most valleys may also be undergoing some kind of downvalley creep or flow. A precise mathematical formulation is problematical; in its simplest form, plastic flow rate $Q$ may be taken to be proportional to the sine of the slope angle (eg. Souchez 1966).

ie.

$$\frac{\partial z}{\partial t} \propto \frac{\partial Q}{\partial x} \text{ or } \frac{\partial z}{\partial t} = C \frac{\partial^2 z}{\partial x^2}$$  \hspace{1cm} (13) \hspace{1cm} C =$ constant

(making the small angle approximation $\sin \alpha \approx \tan \alpha$)

I note here that this equation is the same as equation (1), and is similar to the "landscape-subduing coefficient" of Hirano (1968, 1975). Although equation (13) is similar to the diffusion equation as used by Culling (1963) to describe surface lowering by diffusion creep of soil, my own field observations (Chapter 6) suggest that soil creep processes are much less important than debris slides and slumps in downslope transfer of material on valley side slopes in South
Taranaki. At the simple level of modelling adopted here, equation (13) is adopted to describe the mass movement process.

A composite model of ground surface lowering using equations (11), (12), (13) is

\[ \frac{\partial z}{\partial t} = [A(x+d)e^{-\tau x} \sqrt{1+(\frac{\partial z}{\partial x})^2} + B \sin \theta \sqrt{1+(\frac{\partial z}{\partial x})^2} + C \frac{\partial z}{\partial x} + 2 \sqrt{1+(\frac{\partial z}{\partial x})^2}] \]

(14)

which was simplified by Chappell (1974a) to a more easily manageable linear form

\[ \frac{\partial z}{\partial t} = [A(x+d)e^{-\tau x} + B(x+d)\frac{\partial z}{\partial x} + C \frac{\partial^2 z}{\partial x^2}] \]

(15)

To apply the above equation, four sets of boundary conditions are considered.

(A) Fixed divide at \((x_1, z_1)\) i.e. no headward retreat and fixed slope base at \((x_n, z_n)\) i.e. no basal lowering (see Fig. 7.3B)

To solve equation (15), implicit finite difference methods are used (eg. Hildebrand 1968), and the resultant difference equation is

\[ \frac{1}{K}(z_{i,m+1} - z_i,m) = [A(x_i+d)e^{-\tau x} + B/h(x_i+d) x(z_{i+1,m} - z_{i-1,m} + z_{i+1,m+1} - z_{i-1,m+1})]

+ (z_{i+1,m} + z_{i-1,m} - 2z_i,m + z_{i+1,m+1} + z_{i-1,m+1})] - 2z_i,m+1

\]

ie. collecting like terms

LHS

\[ -z_{i-1,m+1} + \frac{\partial^2 z}{h^2} (x_i+d) - \frac{C}{2h^2} \]

\[ + z_{i,m+1}(1 - \frac{C}{2h^2}) \]

\[ + z_{i+1,m+1} \frac{\partial z}{h} (x_i+d) + \frac{C}{2h^2} \]

RHS

\[ z_{i-1,m+1} \frac{\partial z}{h} (x_i+d) - \frac{C}{2h^2} \]

\[ + z_{i,m} \frac{1}{K} + \frac{C}{2h^2} \]

\[ z_{i+1,m} \frac{1}{h} (x_i+d) + \frac{C}{2h^2} \]

\[ - \frac{C}{2h^2} \]

\[ - \frac{\partial A(x_i+d) e^{-\tau(x_i)}}{h^2} \]
and abbreviating

\[ z_{i-1,m+1}(\alpha_{i-1}) \]

\[ +z_{j,m+1}(\beta) \]

\[ +z_{i+1,m+1}(\lambda_{i+1}) \]

Using the profile at \( t=m \) as boundary condition, the profile at \( t=m+1 \) may be calculated from the following

\[
\begin{pmatrix}
1 & 0 & 0 & 0 & \ldots & z_1 \\
-\alpha_1 & \beta & \lambda_3 & 0 & \ldots & z_2 \\
0 & -\alpha_2 & \beta & \lambda_4 & \ldots & \ddots \\
\vdots & \vdots & \vdots & \ddots & \ddots & \ddots \\
0 & \ldots & \ldots & \ldots & -\alpha_{n-2} & \beta & \lambda_n \\
0 & \ldots & \ldots & \ldots & 0 & 0 & 1 & z_n \\
\end{pmatrix}
\]

\[
= 1 & 0 & 0 & 0 & \ldots & z_1 \\
\alpha_1 & \beta' & -\lambda_3 & 0 & \ldots & z_2 \\
0 & \alpha_2 & \beta' & -\lambda_4 & \ldots & \ddots \\
\vdots & \vdots & \vdots & \ddots & \ddots & \ddots \\
0 & \ldots & \ldots & \ldots & \alpha_{n-2} & \beta' & -\lambda_n \\
0 & \ldots & \ldots & \ldots & 0 & 0 & 1 & z_n \\
\end{pmatrix}
\]

\[ \begin{align*}
\text{(17)} & \\
& \text{ie. } M_1 Z_{m+1} = M_2 Z_m - \Lambda \\
& \text{(18)}
\end{align*}
\]

where \( Z_{m+1}, Z_m, \Lambda \) are vectors; \( M_1, M_2 \) are matrices).
for which the solution is

\[ Z_{m+1} = M_1^{-1} E \]  \hspace{1cm} (19) \]

where vector \( E = M_2 Z_m \) \( \wedge \)

(B) Fixed divide at \((x_k, z_k)\) at time \(j\), with \(k_{j+1} < k_j\) (ie. headward retreat \(x_k(t)\)) and fixed slope base at \((x_n, z_n)\) ie. no basal lowering (Fig. 7.3). In this case \(d(t) = x_k-x_1\) is substituted in equation (17) and expresses uniform headward retreat rate.

(C) Headward retreat as for (B) and basal lowering (Fig. 7.3) with \((x_n, z_n)\) fixed, but \(z_1, z_2, z_3 \ldots \ldots z_{n-1}\) all increased by \(z\) every \(j\) time steps \((j>1)\) ie. uniform downcutting rate. Proceeding as with equation (17).

(D) Variation in rock type ie. erosivity function \((x, z)\) according to position of wave cut surface (Fig. 7.3) relative to ground surface. Takes into account accretion of cover beds on interfluvies at rate \(m\) and basal downcutting at rate \(z\), both relative to wave cut surface. Constants \(A, B, C\) varied according to whether \(z_i\) is above or below the wave cut surface.

To control stability of the difference equation solutions, all calculations are made in a dimensionless form. Equation (17) was solved using a Tectronix 4051 32K mini-computer, using a program provided by Dr J. Chappell.

Best fit curves can be obtained for individual profiles (valley side slopes and stream long profiles) by systematically varying \(A, B, C\) and \(\tau\) until least-squares variance is minimised. For a set of \(n\) profiles,

- (a) least squares variance is minimised as above, and
- (b) the number of model time steps needed for final fits are made similar for profiles of similar age ie. model age variance is minimised.

For a large number of profiles, the above empirical search becomes impracticable, and comparison of the model against a small number of "average" profiles is preferable.
7.3 A DETERMINISTIC MODEL OF VALLEY SIDE SLOPE EVOLUTION IN SOUTH TARANAKI

Initial Conditions: horizontal surface with zero relief (marine/non-marine interface in cover beds)

Boundary conditions: (a) uniform rate of headward retreat
(b) uniform rate of stream downcutting
(c) position of wave cut surface
(d) present valley side slope profile
(e) number of model iterations

Having noted previously (Chapter 6) that a small, but statistically significant, amount of valley side slope asymmetry is present within the terrace front valleys, I have attempted to model east and west facing slopes separately. Seven paired profiles were selected for analysis, each of which exhibits (statistically significant \( p(H_0) < 0.001 \)) a steeper east facing valley side slope (see Table 6.7); the profiles selected were 25, 40, 47, 55, 53/54, 83 and 89.

For east facing valley side slopes a best fit model* is:

\[
\frac{\Delta z}{\Delta t} = 1.35 \times 10^{-3} (x+d) e^{-1.35 \times 10^{-3} x} + 8.12 \times 10^{-4} (x+d) \frac{dz}{dx} + 0.11 \frac{d^2 z}{dx^2} \quad (20)
\]

Goodness of fit is 97% of profile variance explained and 94% of age variance for the set of seven valley side slopes (see Fig. 7.4).

For west facing valley side slopes a best fit model is:

\[
\frac{\Delta z}{\Delta t} = 1.10 \times 10^{-3} (x+d) e^{-1.10 \times 10^{-3} x} + 6.13 \times 10^{-4} (x+d) \frac{dz}{dx} + 0.11 \frac{d^2 z}{dx^2} \quad (21)
\]

In this case goodness of fit is 99% of profile variance explained and 93% of age variance explained for the set of six west facing slopes - see Fig. 7.5 (one slope, 55/2, showed a very much poorer fit (81% of profile variance) and was excluded from the final analysis).

* coefficients reduced by ratio 0.9 below wave cut surfaces.
To interpret the above models I note the following:

(a) The relative weighting of the mass movement factor is increased on west facing slopes relative to east facing slopes.

(b) The values of the constants in equation (20) and (21) do not straightforwardly represent the weightings of the actual processes, since the terms are algebraically different for each process.

Point (a) implies that relative to east facing slopes, mass movement processes are on the average more important as agents of slope development on west facing slopes. The relative importance of the corrasion term is correspondingly reduced on west facing slopes. These relationships suggest that a possible explanation of valley asymmetry in the terrace front valleys lies with the relative importance of mass movements, i.e. more frequent or larger mass movements on west facing slopes. Interestingly, James (1973), in a detailed study of mass movements in the Ruahine Range to the east, has stated that mass movements were most frequently observed on north and east facing slopes (i.e. the opposite trend to that implied by the above model). This apparent discrepancy suggests that a closer analysis of modern processes (particularly of mass movements in relation to aspect) in both areas would be instructive.

Investigation of point (a) above is not a simple matter because the combined action of the three processes is not the sum of their individual actions. However, since the solution term at least is independent of the other two, the interaction of the three processes may not greatly affect such a calculation. Accordingly, by setting the constants, A, B, C to zero as required, the relative weighting of the processes may be approximated by running individual process models and comparing total downcutting over single profiles for each process. In this way it may be estimated that for profile 25/1 (west facing slope), the relative importance of solution : corrasion : mass movement term is approximately 1.6 : 1 : 2.5, while
for profile 25/2 (east facing slope) the relative importance is 1.1 : 1 : 1.6 i.e. approximately 30% of ground lowering is via solution processes.

Few relevant data are available for New Zealand concerning the relative proportion of stream loads carried in solution. Consequently it is difficult to gauge the appropriateness of the above figures. A recent survey by Adams (1979) of stream load data in the North Island indicated that on average 88% of total river load was carried as suspended load, 10% as dissolved load and 2% as bed load, although he pointed out that wide variations occurred. In particular, in catchments where mean annual sediment yields are low, dissolved load may be greater than or equal to suspended load (Adams 1979). In the South Island, where erosion rates are much higher on the average than in the North Island, Adams (1980) reported that suspended load amounts to 93% of total river load, dissolved load 4% and bed load 3%. I conclude therefore, that a 30% solution estimate for surface lowering in the South Taranaki terrace front valleys, is somewhat higher than the New Zealand wide average; however, in an area which is neither seismically as active, nor being uplifted as rapidly, as the eastern north island and Southern Alps of the South Island, such an estimate might not be unreasonable.

The excellent statistical agreement between model and natural slopes (see Fig. 7.4 for model fits), as well as between model time steps and estimated slope ages* would appear to indicate that the model is a more than adequate description of the field data. However, while the solution of equation (17) is unique for specified initial and boundary conditions,

(a) Other mathematical formulations might yield equally well fitting or better fitting models.

* Interpolated assuming uniform headward retreat model.
(b) Different weightings for the constants A, B, C as well as \( \tau \) might yield an equally well fitting model eg. a much more extensive empirical search might yield more than one model of total minimum variance.

With regard to point (a) above I note that Chappell (1974a) produced a statistically close fitting model for stream thalweg evolution in small valleys, incised in coral terraces, which incorporated similar mathematical formulations to those in equation (17). The three processes were weighted in the ratio 1 : 0.3 : 0.7 (solution : corrasion : mass movement) in this model. However, more recently Chappell (1978) has demonstrated that a reformulated model incorporating only a mathematical representation of the effect of solution processes is statistically as good a fit as the previous three-process model. Since the coral terraces are composed almost entirely of CaCO\(_3\), this latter model seems not unreasonable.

I therefore conclude this discussion by emphasising that the models of slope development given above (equations 20 and 21) are two of possibly many mathematical formulations which might yield equally impressive statistical agreement with the field data.

7.4 TOWARDS A DETERMINISTIC MODEL OF STREAM THALWEG EVOLUTION IN SOUTH TARANAKI

Initial conditions: gently sloping surface (marine/non-marine interface in cover beds, truncated by marine cliff.

Boundary conditions: (a) uniform rate of headward retreat
(b) uniform rate of stream downcutting at stream exit
(c) position of wave cut surface
(d) present valley long profile
(e) number of model iterations.

As in the slope modelling exercise above, seven representative valleys were selected for analysis:
SV010, SV018, SV019, SV025 (200Ka age group)
SV024, SV026 (300Ka age group)
SV027 (100Ka age group)

A similar procedure to that adopted for the valley side slope model was followed. As for the previous case, no single process model closely fitted the surveyed long profiles. Furthermore, using the three-process model (as per equation (17)), no adequate fit of all profiles was possible.

This somewhat unexpected result arises because:

(a) The valley long profiles consist of two separate segments.

(i) The valley headwall upstream of the valley-head spring

(ii) The stream thalweg downstream of the valley-head spring.

In particular the model (equation (17)) is unable to adequately model the valley headwall section of the valley long profile.

(b) The valley head spring contributes a significant proportion of streamflow which is derived from subsurface flow immediately inland of the valley head. In other words the model of linearly increasing discharge from the valley head is not a reasonable assumption, and hence the mathematical formulation of the solution and corrision processes in equation (17) is unrealistic.

To incorporate the above observations in the model as it now stands is not a straightforward task, and is beyond the scope of this investigation. For example, it is required that we know

(1) the proportion of total stream discharge originating at the valley head spring or as an approximation of this,

(2) the horizontal length of the subsurface catchment area ie. the position of the groundwater divide,

as well as

(3) specific information regarding the mechanics and geometry of valley head retreat eg. does the headwall height remain roughly constant or if not, how does it vary with progressive retreat.
As a first approximation, assume that approximately 20% of baseflow stream discharge originates as subsurface flow (roughly in accord with discharge measurements made in SV001 and qualitative estimates in other surveyed valleys). In addition assume a linear increase in discharge from the groundwater catchment divide; so that the position of the base of the headwall \( x_h \) varies such that \( x_h = x_1 - (x+d)/4 \) for constant headward retreat and fixed subsurface flow distance \( f = (x+d)/4 \) (see Fig. 7.6A).

Using this model, a best fit model for the seven valley long profiles is

\[
\frac{\Delta z}{\Delta t} = 1.37 \times 10^{-4} (x+d) e^{-7.1 \times 10^{-4} x} + 8.4 \times 10^{-5} (x+d) \frac{dz}{dt} + 0.1 \frac{d^2 z}{dx^2} \tag{22}
\]

However, even this model is a poor fit for SV010 and SV027. For the other 5 profiles goodness of fit is 95% of profile variance explained downstream of the valley head spring and 90% of age variance explained (see Fig. 7.6B for model fits). I note that SV010 and SV027 have two of the highest implied headward retreat rates of the 27 surveyed terrace front valleys (Table 6.1). This suggests that one possible explanation of the poor fit is that they did not evolve by a process of headward retreat, but originated as terrace surface drainage soon after terrace emergence. An additional source of uncertainty is the estimate of 20% subsurface discharge.

For the set of valleys as a whole the above model (equation (22)) yields process weightings solution: corrasion : mass movement in the proportions 30% : 30% : 40%. Although these weightings are similar to those for the slope models developed previously, intuitively one might expect that stream thalweg lowering would be dominated by corrasion processes since

(a) Baseflow is a major component of stream flow, and conductivity measurements indicate little change in dissolved ion content downstream of valley head springs.
(b) Mass movement processes appear to be most important at the valley headwall, and contribute little to overall sediment movement downstream of headwater tracts.

I conclude therefore that the modelling of stream thalweg evolution is not satisfactorily achieved by the above model.

As a final conclusion, in respect of the modelling of both valley side slopes and stream thalwegs, it is important to remember that the statistical curve fitting exercise attempted here incorporates several variables which are adjusted until variance is minimised. In the same way that a polynomial function of degree n can always be found to fit n+1 data points, so for cases where the number of data points is small and the number of variables is large, little meaning can be attached to models explaining a high proportion of profile variance.
CHAPTER 8

CONCLUSIONS AND DIRECTIONS FOR FUTURE RESEARCH

This thesis is composed of two major parts: the first is concerned with stratigraphy and dating of the South Taranaki terraces, and the second relates to their geomorphic evolution. The latter relies heavily on the former, although the converse is not true.

On reflection, the geological portion of the thesis is seen as the more satisfying piece of work, not only because the conclusions are more soundly based, but also because they have immediate implications: Quaternary continental records are fragmentary compared to the more or less continuous depositional records in deep sea sediments. This has been particularly evident in New Zealand, where existing climatic stratigraphy has sought to link glacial deposits in the South Island with interglacial marine deposits in the North Island. Such correlations have been hindered by a lack of chronological control in both areas. The stratigraphy of the South Taranaki terrace sequence has a more complete, and better dated record than that upon which the existing New Zealand Upper Quaternary Stages were founded (see Table 5.13). Apart from the relative completeness of the South Taranaki stratigraphic record, there are other obvious advantages in having type sections within one area eg. the possibility of miscorrelation between sections is reduced accordingly. The South Taranaki marine terrace sequence would therefore provide a much better basis for Upper Quaternary chronostratigraphic subdivision, than the existing climato-stratigraphic subdivision.

On the other hand the stratigraphy of the terrace sequence is complex, and at best this study is a reconnaissance mission, upon which subsequent workers can build. Chronology at present rests upon
a single fission track date (Omahina Tephra) and on thirty or so amino acid recemisation dates, which are in turn tied to the fission track age estimate. Future workers, particularly in the Wanganui area, will undoubtedly discover other rhyolitic tephras suitable for fission track dating. In particular the unequivocal relation of the Mt Curl Tephra to the terraces remains to be unravelled, while the possibility of discovering the Rewa Pumice and/or other tephras within the cover beds of the oldest terraces is an exciting prospect. Other dating methods such as U/th dating, $^{14}C$ accelerator dating and thermoluminescence dating could usefully be applied to materials within the terrace cover beds.

This study has without question demonstrated the potential usefulness of the amino acid racemisation dating technique to Quaternary stratigraphic studies. The widespread occurrence of fossil woods within the terrace sequence has enabled age estimates of several of the terraces to be made. The South Taranaki terraces provide an ideal setting in which to further test the dating method because:

(a) The area is climatically fairly uniform (cf. dependence of racemisation rate constants on temperature).

(b) Multiple samples may be collected from similar stratigraphic positions over a wide area.

(c) Relatively high sedimentation rates increase the likelihood of preservation and reduce the effects of soil temperature variations which may occur in near surface sites.

(d) Cross checking with other dating methods will be possible.

Rates and styles of Upper Quaternary deformation have been deduced for the South Taranaki terrace sequence. Calculated mean uplift rates vary between 20 and 70cm/Ka; while rates of tilting vary between 7 and 27nrad/year. Similar rates of deformation have been reported from other uplifted marine terrace sequences in areas adjacent to actively converging plate boundaries overseas (California, Japan, New Guinea, Indonesia, Loyalty Islands), as well as elsewhere within New Zealand (eg. Yoshikawa et al 1980, Ghani 1978). Further
work is clearly warranted to focus more closely on tectonic relationships between the South Taranaki marine terraces and the South Wanganui Basin. An implication of this study is that the terrace strandlines older than the Brunswick terraces represent the shoreline equivalents of Upper Castlecliffian marine sediments within the Wanganui Basin. Terrace deformation is therefore closely linked with the final uplift and death of the South Wanganui Basin.

The sediments of the South Wanganui Basin contain an even more detailed record of sea level fluctuations than the terraces themselves (cf. diagram in Fleming 1953, p.302). In particular, a record of relative low sea level events is present, which with careful surveying, and paleontological work, would provide one of the best relative sea level curves available, for the critical, as yet poorly dated portion of the Upper Quaternary beyond c.200Ka BP. Furthermore the presence of many rhyolitic tephras (both dated and undated) within the basin sequence, makes it chronologically an excellent proposition. Micropaleontological work will enable detailed biostratigraphic correlations to be made between the Wanganui Basin shallow marine sediments, and those of deep sea cores. A combination of tephro-chronology and biostratigraphy will also enable correlation with the important Quaternary shallow marine sequence near Cape Kidnappers in Hawke Bay (eastern North Island, New Zealand - cf. Kamp (1979), and continuing work (P. Kamp pers comm 1980).

Recommended future studies on the South Taranaki terraces may also include:

(a) Palynological investigation of carbonaceous deposits within the terrace cover beds - a detailed vegetation record of this kind which is stratigraphically tied to Quaternary marine terraces has yet to be described from anywhere else in the world to my knowledge.

(b) Provenance studies of the marine sands within the terraces, would shed considerable light on the volcanic history of western Taranaki. In addition such studies could yield quantitative information on rates and styles of mineral diagenesis.
(c) Quantitative study of weathering rates and pedogenesis using dated deposits of similar composition (eg. tephras or loess) as a chronosequence. Such a study would not only yield interesting new data on long term weathering rates and mechanisms, but could also contribute to our understanding of the evolution of the South Taranaki landscape.

Quite clearly the chronology developed in this study enables quantitative long term rates of geologic and geomorphic evolution to be calculated. In particular, terrace sequences of this kind are amenable to the method of reconstruction of initial forms, and as such are particularly valuable to geomorphologists. The reconstruction and testing of quantitative evolutionary models as attempted here, reveals the complexity of evolutionary changes even in an apparently simple field situation. Similar studies by Chappell (1974a) and Dunkerley (1977a, b) confirm this point, although on the Huon Peninsula coral terraces, which both these workers studied, solution processes appear to be the dominant mode of landscape change; such single process landscapes offer a better prospect at this stage of our knowledge, for understanding the way landscapes evolve in terms of quantitative assessment of processes. On the other hand, the multi-process landscape of the South Taranaki terraces contains a record of its vegetation history: Vegetation changes from lowland podocarp forest to open scrub/grassland such as have been identified at the Inaha type section (Fig. 4.10) by McGlone et al (1979) must exert strong influences on the relative intensity and interaction of geomorphic processes. Without such records of vegetation change, one must assume that process mechanisms and interactions remain sufficiently unaltered to allow mathematical deterministic models of landscape development to model at least the long term "average.

It is evident at this stage, that the geomorphic data and models described in this thesis now require a programme of field measurement of processes for their further development. If nothing else they serve to more closely identify the field data required to
understand long term landscape development. To illustrate these points, the models of slope development suggest that solution, corrosion and mass movement processes contribute to ground surface lowering in the proportions 30% : 20% : 50%; to test this assertion, the following measurements might be carried out:

(a) Measurement of solution processes in selected small catchments containing a large proportion of their natural vegetation. This would involve multiple down valley measurements of conductivity and dissolved ion concentrations during baseflow and stormflow, as well as spring water sampling and runoff plots on valley side slopes. Such a program could conceivably be managed within a period of a year, and provide fair estimates of dissolved load concentrations, composition and solute dynamics within terrace front valleys.

(b) Suspended load sampling could be carried out at the same time as the above measurements, although the significance of high magnitude/low frequency events (e.g. storms, earthquake-induced slope failure) in supplying suspended load could not be established in such a short period.

(c) Effective direct measurement of ground surface lowering is difficult, principally because of its slowness relative to the period of observation. In addition, although certain mass movements such as debris slides and rotational slumps produce more rapid and noticeable changes, their intermittent nature and often indeterminate place of occurrence, make instrumentation procedures ineffective. Aerial photography interpretation and field checking over a number of years might therefore be the best course of action with regard to mass movement processes.

I conclude this thesis in the hope that the data presented here may act as a stimulus to future geologic and geomorphic research on the South Taranaki terrace sequence, in particular towards our understanding of the long term evolution of landscapes.
Plate 2.1: Eastern side of Mt Egmont (main cone) and Fanthams Peak (parasitic cone). Vegetation grades upwards from lowland podocarp-hardwood forest to upland podocarp-hardwood forest to subalpine scrub to alpine grassland. May 1980 photograph.

Plate 2.2: Egmont (left), Pouakai (centre) and Kaitake (right) showing progressive dissection with increasing age and southerly migration of volcanic activity.
Plate 2.3: Dissection of Pliocene shallow marine sediments inland of the terrace sequence near grid reference N129/090222.

Plate 2.4: Summit accordance rising inland towards Mt Ruapehu (faintly visible centre-left skyline). Brunswick Terrace surface in foreground near N130/238124.
Plate 2.5: Aerial view of Brunswick Terrace dissection with wide, flat interfluves in foreground. Kaiatea terraces plainly visible further inland near Hurleyville (N129/058213).

Plate 2.6: Whenuakura River valley looking inland from N136/075055. Kaiatea Terraces are on the skyline.
Plate 2.7: Talus cones on coastal cliff east of Ohawe Beach near N129/794268. Rapanui wave cut surface and cover beds exposed in cliff face.

Plate 2.8: Brunswick wave cut surface and cover beds exposed in valley side near N137/297077.
Plate 2.9: Ngarino cliff (surface expression) inland from Waverley near N130/195127 looking east.

Plate 2.10: Steam dissection of Ngarino cliff inland from Waverley near N130/193127 looking north.
Plate 2.11: Trough cross-stratified, andesitic marine sands, Inaha Terrace near NI29/748278 (type section).

Plate 2.12: Small scale cross-stratified and laminated, andesitic marine sands, Kaiatea III Terrace near NI30/184160.
Plate 2.13: Inland edge of active coastal dunes west of Okehu Stream near N137/415941.
Plate 4.1: Omahina Tephra (white, rhyolitic, 10cm thick) within dark brown, weathered andesitic ashes. Ball Road Extension near N129/087238.
Plate 4.2: Omahina Tephra at type section on Omahina Road (N130/237150). Dark slot in base of tephra (right-centre) is position from which the sample for fission-track dating was obtained. Note dark olive-brown, nutty structured clay above tephra. Tape is 2m long.
Plate 4.3: Rangitawa Pumice type section, north bank of Rangitawa Stream (N143/962624). Head of geological hammer (30cm high) rests on base of Rangitawa Pumice in centre of photograph. See also Fig. 4.1.

Plate 4.4: Zoned siltstone clasts forming basal conglomerate, Kaiatea I Formation. Type section on Okahutiria Road (N130/187187). Laminated marine sands overlie the conglomerate. Wave cut surface obscured by slumping near base of exposure.
Plate 4.5: Quarry near Brunswick Road at N138/582940. Proposed new type section for Terangian Stage. Fossiliferous, laminated marine sands exposed in pit face. Brunswick Terrace. See also Fig. 4.4.

Plate 4.6: Upper Brunswick Terrace (centre distance) looking west from Omahina Road (N130/225127) towards Braemore Road (visible centre-left, in distance, on Upper Brunswick cliff line). Surface expression of Brunswick cliff, truncating Upper Brunswick Terrace, clearly visible in middle distance to right of telegraph poles. Brunswick Terrace surface in foreground.
Plate 4.7: Ngarino Terrace cover beds exposed on Omahina Road at N130/217108. Note prominent orange-brown, weathered ash. See also Fig. 4.7.

Plate 4.8: Rapanui Terrace cover beds exposed on Kohi Road at N137/150088. Note thin orange-brown weathered ash. Grey, pebbly, marine sands visible in lowest portion of cutting. See also Fig. 4.8.
Plate 4.9: Rapanui/Inaha cliff circa 1.5km east of Manawapou River mouth at N129/927185.
Plate 4.10: Rapanui cover beds east of Ohawe Beach at mouth of unnamed stream (N129/793269). See also Fig. 4.8.
Plate 4.11: Inaha Terrace cover beds at type section N129/748278. See also Fig. 4.10.
Plate 4.12: Aerial view of mouth of Waingongoro River (N129/772277). Rapanui wave cut surface exposed in coastal cliffs in right foreground. Inaha Terrace surface and cliff on western (left) side of river mouth.

Plate 4.13: Tree stumps in growth position at Waverley Beach (N137/185985), within unnamed post-Inaha marine terrace deposits. See also Fig. 4.11.
Plate 6.1: Progressive dissection of Kaiatea Terraces inland of Hawera. SV001 (left valley) and SV002 (right valley) are incised into Kaiatea III Terrace in foreground. Feral (fine textured relief) is visible in distance.

Plate 6.2: Contrasting geomorphic responses above and below a wave cut surface (arrowed) overlying Pliocene mudrocks, North Taranaki.
Plate 6.3: A small shore-parallel valley inland from Hawera looking east. Wave cut surface dips from left to right. Note prominent spring line (vegetation line) and spur and groove topography with seepage heads on updip (left) valley side; relatively smooth, steeper downdip (right) valley side lacking springs.

Plate 6.4: Updip valley side slope of shore-parallel valley above; well developed seepage heads associated with wave cut surface spring line (arrowed); prominent spur and groove topography.
Plate 6.5: Small amphitheatre shaped seepage head with steep, terraced headwall, and swampy floor; vegetation line marks Kaiatea III wave-cut surface.

Plate 6.6: Seepage heads and mass-movement scars above Kaiatea III wave-cut surface (vegetation line). Well developed spur and groove topography.
Plate 6.7: Amphitheatre basin head, SV001 (see also Plate 6.1). Headwall spring (arrowed) and several auxiliary springs clearly visible. A shallow, gently sloping, basin is present upstream of the headwall. Note lack of defined channel on swampy valley floor.

Plate 6.8: SV001, looking upstream; vertical relief c.40m; rectilinear to slightly convex valley side slopes.
Plate 6.9: SV015 (shore-normal valley) looking upstream from outer edge of Brunswick Terrace. Note well developed spur and groove topography, terracettes, mass movement scars and infilled valley floor; simple valley side slopes consisting of two essentially rectilinear segments above (gentler gradient) and below (steeper gradient) the wave cut surface (arrowed) are clearly evident on right hand valley side.

Plate 6.10: Valley floor SV006 (shore-normal valley).
Plate 6.11: Large debris slide in Brunswick Terrace cover beds; upper limit of vegetation line marks wave cut surface.

Plate 6.12: Rotational slumps with basal earthflows, Brunswick (?) Terrace cover beds east of Wanganui.
South Taranaki-Wanganui marine terraces
South Wanganui Basin
Taupo Volcanic Zone
Alpine Fault
Plate boundaries

Fig. 2.3 Tectonic Setting
Relative plate movement (large arrows) in mm/year from Walcott (1978)
Bathymetric contours in metres
Fig. 2.5 Terrace Morphology and Nomenclature
Stage 1: Rapid sea level rise $\gg$ uplift rate. Little platform cutting
Stage 2: Slow sea level rise $>$ uplift rate. Platform cutting important; truncation of previous terrace; transgressive gravels deposited
Stage 3: Sea level constant with respect to uplift. Completion of terrace cutting; strandline active
Stage 4: Slow sea level fall opposed to uplift. Regressive marine and beach sands deposited on terrace
Stage 5: Rapid sea level fall opposed to uplift. Emergence without cliffing

Fig. 2.6 Sea Level and Terrace Development

 Modified after Bradley and Griggs (1976)
Fig. 2.7 Climatic Data
Fig. 2.8 Lapse Rate

Data from NZ Meteorological Service (1978)
Fig. 2.9 Vegetation

Modified from Wendelken (1976), Nicholls (1974)
Fig. 3.1(a) Spatial representation of the D- and L-enantiomers of \( \alpha \)-amino carboxylic acids. Mechanism of racemisation involves formation of a carbanion intermediate.

Fig. 3.1(b) Stoichiometric formulas of four theoretically possible isomers of isoleucine.
(A) Protein Amino Acids

<table>
<thead>
<tr>
<th>Alanine</th>
<th>Arginine</th>
<th>Aspartic Acid</th>
<th>Asparagine</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH₃-C-COOH</td>
<td>H₂N-C-NH-CH₂-CH₂-CH₂-C-COOH</td>
<td>H₂O-C-CH₂-C-COOH</td>
<td>H₃N-C-CH₂-C-COOH</td>
</tr>
<tr>
<td>NH₂</td>
<td>NH</td>
<td>NH₂</td>
<td>NH₂</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Cysteine</th>
<th>Glutamic Acid</th>
<th>Glutamine</th>
<th>Glycine</th>
</tr>
</thead>
<tbody>
<tr>
<td>HS-CH₂-C-COOH</td>
<td>HO-C-CH₂-CH₂-C-COOH</td>
<td>H₂N-C-CH₂-CH₂-CH₂-C-COOH</td>
<td>H₂N-CH₂-C-COOH</td>
</tr>
<tr>
<td>NH₂</td>
<td>NH</td>
<td>NH₂</td>
<td>NH₂</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Histidine</th>
<th>Isoleucine</th>
<th>Leucine</th>
<th>Lysine</th>
</tr>
</thead>
<tbody>
<tr>
<td>H₂C=CH₂-C-COOH</td>
<td>CH₃-C-C-C-COOH</td>
<td>CH₃-C-OH-C-COOH</td>
<td>H₂N-CH₂-C-CH₂-CH₂-C-COOH</td>
</tr>
<tr>
<td>NH</td>
<td>NH₂</td>
<td>CH₃</td>
<td>NH₂</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Methionine</th>
<th>Phenylalanine</th>
<th>Proline</th>
<th>Serine</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH₃-S-CH₂-C-COOH</td>
<td>C₆H₄-C-C-C-COOH</td>
<td>H₂O-C-CH₂-C-COOH</td>
<td>H₂O-C-C-C-COOH</td>
</tr>
<tr>
<td>NH₂</td>
<td>NH₂</td>
<td>NH</td>
<td>NH₂</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Threonine</th>
<th>Tryptophan</th>
<th>Tyrosine</th>
<th>Valine</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH₃-C-C-C-COOH</td>
<td>C₆H₄-C-C-C-C-C-COOH</td>
<td>HO-CH₂-C-C-C-COOH</td>
<td>CH₃-C-C-C-C-C-C-COOH</td>
</tr>
<tr>
<td>OH</td>
<td>NH₂</td>
<td>NH²</td>
<td>NH₂</td>
</tr>
</tbody>
</table>

(B) Collagen Amino Acids

<table>
<thead>
<tr>
<th>4-Hydroxyproline</th>
<th>5-Hydroxysine</th>
</tr>
</thead>
<tbody>
<tr>
<td>HO-CH₂-C-C-C-COOH</td>
<td>H₂N-CH₂-C-CH₂-CH₂-C-COOH</td>
</tr>
<tr>
<td>NH</td>
<td>OH</td>
</tr>
</tbody>
</table>

(C) Non-Protein Amino Acids

<table>
<thead>
<tr>
<th>α-Aminobutyric Acid</th>
<th>β-Alanine</th>
<th>7-Aminobutyric Acid</th>
<th>Ornithine</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH₃-C-C-C-COOH</td>
<td>HOOC-CH₂-C-COOH</td>
<td>HOOC-CH₂-C-C-C-COOH</td>
<td>H₂N-CH₂-C-CH₂-CH₂-C-COOH</td>
</tr>
<tr>
<td>NH₂</td>
<td>NH</td>
<td>NH₂</td>
<td>NH₂</td>
</tr>
</tbody>
</table>

Fig. 3.2 Stoichiometric formulas of twenty six naturally occurring amino acids.

(C*: asymmetric carbon atom) Glycine, β-alanine and 7-aminobutyric acid have no centre of optical asymmetry; isoleucine, threonine, hydroxyproline and hydroxysine have 2 centres; all others have a single asymmetric centre (at the α-carbon position).
Fig. 3.3(a) Protein hydrolysis.

Fig. 3.3(b) Chromatogram (spectrum) of a synthetic amino acid using the modified Technicon Auto Analyser (modified after James, 1972).
Fig. 3.4 Arrhenius Plot: Isoleucine—Alloisoleucine Epimerisation Kinetics
Fig. 3.5 Contamination sources (losses & additions) for fossil proteins
Fig. 3.6 Effect of equilibrium constant ($K_{eq}$) on model ages.
Figs. 3.7-3.10 Changes in amino acid composition with time (D/L ratio)
Fig. 3.11 South Taranaki Dating Curve

Point A: initial D/L = 0.01
Point B: 400 Ka calibration
mean D/L = 0.112
K_s = 2.59 \times 10^{-7} yr^{-1}

C^{14} age 33,000 ± 1100 (ANU-1887)
C^{14} age 7500 ± 110 (ANU-1888)
C^{14} age 6200 ± 100 (ANU-1891)
C^{14} age 22,700 ± 600 (NZ-16238)

Sample numbers refer to those with BIP prefix in Table 3.4
A Crude temperature model for past c.400 Ka, loosely based on sea level curves and δ¹⁸O curves shown in Fig. 5.4.

B Glacial-Interglacial temperature difference calculated using South Taranaki data at sea level.

3.12 Paleotemperature Models
Fig. 3.13 Model Errors: Summary.
Fig. 4.2 Kaiatea and Pre-Kaiatea Terraces: Map and Sections.
Fig. 4.3 Brunswick Terraces: Map and Sections.
Fig. 4.4 Brunswick Type Section N137/502981.
Fig. 4.5 Brunswick Terrace: Warrengate Section N138/732815.
Fig. 4.6 Ngarino Terrace: Map and Sections.
Fig. 4.7 Rapanui Terrace: Map and Sections.
Ohawe Beach Section
N129/750278 to N129/753269
(horizontal distance not to scale)

Fig. 4.8 Inaha Terrace: Map and Sections.
Fig. 4.9 Inaha Type Section N129/748278.
Fig. 4.10 Younger Deposits: Map and Sections; Location of C¹⁴ Dated Samples.
Fig. 5.1 Elementary Deformation Patterns
Fig. 5.3 Relation Diagrams and Shore Normal Deformation Models

(a) Constant uplift on any single shore normal traverse, but varying between traverses (T1, T2, T3, T4).

(b) Constant bending type uplift on any single shore normal traverse, but mean rate varying between traverses (T1, T2, T3, T4).

(c) Dome-type uplift on any single shore normal traverse, but varying between traverses (T1, T2, T3, T4).

(d) Traverse strike on tilting surface for models (a), (b), and (c).
Fig. 5.4 Sea Level Chronology

A. $\delta^{18}O$ variations in Caribbean core V12-122 (Broeker and Van Donk, 1970); glacial terminations numbered; position of U/V paleontologic boundary indicated

B. $\delta^{18}O$ variations in equatorial Pacific core V28-238 (Shackleton and Opdyke, 1973); odd numbered (warm) stages indicated

C. Sea level variations identified in New Guinea and Atauro (Chappell, 1981); * dated reef crests in Barbados (Bender, et al, 1979)

D. Strandline ages, South Taranaki (this work)
Fig. 5.5 Uplift Rates (mm/yr)/Terrace Deformation

- 0.65 mm/yr
- 0.60 mm/yr
- 0.55 mm/yr
- 0.50 mm/yr
- 0.45 mm/yr
- 0.40 mm/yr
- 0.35 mm/yr
- 0.30 mm/yr

- Uplift rate Kaiatea Strandline
- Uplift rate Rapanui Strandline
- Uplift rate other strandlines (hinging uplift model)
Fig. 5.6 Relation Diagram

A: terrace remnant Mowhanau Stream mouth (N137/440917)
B: terrace remnant Tangahoe/Manawapo River mouths (N129/915196)
Fig. 5.7 Summary

A. Dated rhyolitic tephras, South Taranaki
B. Modified N.Z. stages (Oturian = Rapanui Terrace, Terangian = Brunswick Terrace)
C. Hawera Series/Castlecliffian Stage boundary (Boellstorff and Te Punga, 1977)
D. Modified N.Z. stages (Oturian = last interglacial, Terangian = penultimate interglacial)
E. Dated lignites, South Taranaki
F. Sea level variations, South Taranaki
   E = Estuarine sediments overlying marine sediments
   T = Minimum estimate of sea level fall
G. $\delta^{18}O$ variations in equatorial Pacific core V28-238 (Shackleton and Opdyke, 1973); odd numbered (warm) stages indicated
Fig. 6.1 Terrace Data, Waverley Area.

A. Percent undissected terrace surface remaining (interfluve area)

\[ Y = 100 - 0.19X \]
\( r^2 = 0.94 \)

B. Drainage density (km/\( \text{km}^2 \))

\[ Y = 0.005X \]
\( r^2 = 0.94 \)

C. Mean non-marine cover bed thickness (m)

\[ Y = 0.06X \]
\( r^2 = 0.96 \)
Fig. 6.2 Drainage Development.

A: Regression

B: Transgression

C: Small streams

D: Small streams

E: Larger streams

F: Larger streams

G: Stable sea level
Fig. 6.3 Typical Terrace Front Valley.
Fig. 6.4 Surveyed Thalwegs, Terrace Front Valleys.
Fig. 6.5 Stream Thalwegs
(examples of power curve fits)
Fig. 6.6 Slope Data.

(a) Histogram

Means
26.8° 28.2° 31.4°

All slopes  n = 5060
Cover beds  n = 3357 (66%)
Mudstone  n = 1703 (34%)

(b) Sample Profiles
(scaled to equal length)

29/3

60

47/1

50/2

wave cut surface

stream thalweg

Fig. 6.6 Slope Data.
Fig 6.7
Fig. 6.8 Orientation of Slope Profiles (10° Sectors) Relative to Magnetic North
Frame of reference: stream long profile

valley head

valley depth (D)

terrace surface

distance downstream (L)

stream thalweg

D = \alpha L^B

A  
No headward retreat; position of valley head fixed

\[ r_{di} < r_{dj} < r_{dk} < r_{dl} \]

B  
Constant headward retreat rate

\[ r_{di} > r_{dj} > r_{dk} > r_{dl} \]

C  
Accelerating headward retreat rate

\[ r_{di} > r_{dj} > r_{dk} > r_{dl} \]

D  
Decelerating headward retreat rate

\[ r_{di} \neq r_{dj} \neq r_{dk} \neq r_{dl} \]

Fig 7.1  
Models of Stream Downcutting & Headward Retreat
Frame of reference: rectilinear valley side slope

Divide

terrace surface

distance downslope (L)

valley side slope

$D = aL$

valley depth (D)

A Fixed divide, no valley widening

slope age (t)

downcutting rate ($r_g$)

$r_{gi} < r_{gj} < r_{gk} < r_{gl}$

B Constant valley widening

C Accelerating valley widening

D Decelerating valley widening

$r_{gi} < r_{gj} < r_{gk} < r_{gl}$

Fig 7.2 Models of Stream Downcutting & Valley Side Slope Development
A GROUNDSURFACE LOWERING

B NO HEADWARD RETREAT: NO BASAL LOWERING
fixed divide/fixed critical distance (d)

C HEADWARD RETREAT: BASAL LOWERING
fixed divide/critical distance varies $x_k(t)$

D WAVECUT SURFACE POSITION
wavecut surface position varies relative to $z_n$ (basal lowering) and $z_1$ (cover bed accretion) according to $z_m(t)$

Fig 7.3 Model Frameworks
Fig 7.4 East Facing Slopes: Model Fits
Fig 7.5 West Facing Slopes : Model Fits
Fig. 7.6 Valley Long Profiles
TABLE 3.1: VARIATION OF HALF LIFE WITH TEMPERATURE: ISOLEUCINE IN FOSSIL WOOD

Using $K_{ISO} = A \exp \left( -\frac{E}{RT} \right)$

where $E = 29400 \text{ cal.mol}^{-1}$

$A = 3.853 \times 10^{12} \text{ hr}^{-1}$

$R = 1.987 \text{ cal.deg}^{-1} \text{ mol}^{-1}$

$K' = 1/Keq = 1/0.86 = 1.163$

<table>
<thead>
<tr>
<th>$T(\text{°K})$</th>
<th>$K_{ISO}$ (yr$^{-1}$)</th>
<th>$\tau = \ln \frac{2}{(1+K')} \cdot K_{ISO}$ (yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>273</td>
<td>$9.78 \times 10^{-8}$</td>
<td>3,280,000</td>
</tr>
<tr>
<td>280</td>
<td>$3.79 \times 10^{-7}$</td>
<td>845,000</td>
</tr>
<tr>
<td>281</td>
<td>$4.69 \times 10^{-7}$</td>
<td>683,000</td>
</tr>
<tr>
<td>282</td>
<td>$5.52 \times 10^{-7}$</td>
<td>581,000</td>
</tr>
<tr>
<td>283</td>
<td>$6.64 \times 10^{-7}$</td>
<td>483,000</td>
</tr>
<tr>
<td>288</td>
<td>$1.65 \times 10^{-7}$</td>
<td>194,000</td>
</tr>
<tr>
<td>293</td>
<td>$3.96 \times 10^{-6}$</td>
<td>80,900</td>
</tr>
<tr>
<td>298</td>
<td>$9.23 \times 10^{-6}$</td>
<td>34,700</td>
</tr>
<tr>
<td>373</td>
<td>$2.00 \times 10^{-1}$</td>
<td>1.60</td>
</tr>
</tbody>
</table>

NB. $D/L_T = 0.30$
<table>
<thead>
<tr>
<th>Factor</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. TEMPERATURE</td>
<td>described by Arrhenius equation ( K_L = A \exp(-E/RT) )</td>
</tr>
<tr>
<td>2. MATRIX MATERIAL</td>
<td>eg. bone, carbonate, wood, silica</td>
</tr>
<tr>
<td>3. FOSSIL SPECIES</td>
<td>eg. foraminifera species, mollusc species</td>
</tr>
<tr>
<td>4. HYDROLYSIS STATE</td>
<td>whether free, peptide or protein bound</td>
</tr>
<tr>
<td>5. CLAY MINERALS</td>
<td>catalysis via surface properties</td>
</tr>
<tr>
<td>6. HYDRATION</td>
<td>effects uncertain; at least some ( H_2O ) required for racemisation to proceed</td>
</tr>
<tr>
<td>7. SITE IN PROTEIN CHAIN</td>
<td>effect of adjacent amino acids</td>
</tr>
<tr>
<td>8. pH</td>
<td>little effect in range 5 to 8</td>
</tr>
<tr>
<td>9. IONIC STRENGTH</td>
<td>effect uncertain</td>
</tr>
</tbody>
</table>

* Rate of racemisation varies from amino acid to amino acid, apparently according to electron-withdrawing capacity of adjacent R-groups on α-carbon atom.
### TABLE 3.3: $^{14}C$ DATES FOR SOUTH TARANAKI WOOD SAMPLES

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Location</th>
<th>NZ/ANU Lab. No.</th>
<th>$^{14}C$ Age</th>
<th>Bound D/L ratio</th>
<th>Racemisation age</th>
</tr>
</thead>
<tbody>
<tr>
<td>BJP-001</td>
<td>N129/786289</td>
<td>ANU-1880</td>
<td>&gt;35,200*</td>
<td>.029*</td>
<td>74,000</td>
</tr>
<tr>
<td>BJP-002</td>
<td>N129/769277</td>
<td>ANU-1881</td>
<td>Background</td>
<td>&lt;.045*</td>
<td>&lt;136,000</td>
</tr>
<tr>
<td>BJP-004</td>
<td>N129/918195</td>
<td>ANU-1882</td>
<td>23,100±500</td>
<td>&gt;.012*</td>
<td>&gt;8,000</td>
</tr>
<tr>
<td>BJP-006</td>
<td>N129/918195</td>
<td>ANU-1884</td>
<td>37,900±1800</td>
<td>&gt;.022*</td>
<td>&gt;46,000</td>
</tr>
<tr>
<td>BJP-007</td>
<td>N129/773276</td>
<td>ANU-1885</td>
<td>&gt;35,200*</td>
<td>nd</td>
<td>-</td>
</tr>
<tr>
<td>BJP-008</td>
<td>N129/917196</td>
<td>ANU-1886</td>
<td>6300±100</td>
<td>.016</td>
<td>23,000</td>
</tr>
<tr>
<td>BJP-015</td>
<td>N129/748278</td>
<td>ANU-1887</td>
<td>33,300±1100</td>
<td>.014</td>
<td>15,000</td>
</tr>
<tr>
<td>BJP-037</td>
<td>N137/192983</td>
<td>ANU-1888</td>
<td>7500±110</td>
<td>nd</td>
<td>-</td>
</tr>
<tr>
<td>BJP-039</td>
<td>N137/185985</td>
<td>ANU-1890</td>
<td>7300±110</td>
<td>.017</td>
<td>27,000</td>
</tr>
<tr>
<td>BJP-040</td>
<td>N137/195985</td>
<td>ANU-1891</td>
<td>6200±100</td>
<td>nd</td>
<td>-</td>
</tr>
<tr>
<td>BJP-052</td>
<td>N129/918194</td>
<td>ANU-1892</td>
<td>8100±110</td>
<td>nd</td>
<td>-</td>
</tr>
<tr>
<td>BJP-046</td>
<td>N108/375714</td>
<td>NZ13619</td>
<td>18,900±500</td>
<td>.012</td>
<td>8,000</td>
</tr>
<tr>
<td>BJP-048</td>
<td>N118/403473</td>
<td>NZ1623B</td>
<td>22,700±600</td>
<td>.030</td>
<td>78,000</td>
</tr>
<tr>
<td>BJP-033</td>
<td>N129/877261</td>
<td>NZ3970B</td>
<td>&gt;43,900</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

+ within 1 standard deviation of background

1. D/L ratio (L-isoleucine/D-alloisoleucine) of amino acids presumed to be entirely within residual proteins after pretreatment with 6NHCl for 24 hours at room temperate.

2. calibrated to BJP-026, 027, 028, 029 (400Ka BP) : see Table 3.4

* interference from overlapping chromatogram peaks.
### Table 3.4: Racemisation Dates Calibrated to 400ka Meremere Lignite

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Location/Description</th>
<th>Bound$^{D/L}$</th>
<th>Age (Ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BJP-001X</td>
<td>N129/786289: River terrace incised below Rapanui wave cut surface</td>
<td>0.029</td>
<td>74</td>
</tr>
<tr>
<td>BJP-002X</td>
<td>N129/769277: Fluvial deposit incised below Inaha wave cut surface</td>
<td>0.045</td>
<td>&gt;130</td>
</tr>
<tr>
<td>BJP-004X</td>
<td>N129/918195: River terrace near HM</td>
<td>&gt;0.012</td>
<td>&gt;3</td>
</tr>
<tr>
<td>BJP-005X</td>
<td>N129/918195: River terrace near HM</td>
<td>&gt;0.022</td>
<td>&gt;46</td>
</tr>
<tr>
<td>BJP-010X</td>
<td>N129/748278: Inaha terrace (Mania Lignite)</td>
<td>0.034</td>
<td>93</td>
</tr>
<tr>
<td>BJP-014X</td>
<td>N129/748278: Inaha terrace</td>
<td>0.018</td>
<td>31</td>
</tr>
<tr>
<td>BJP-015X</td>
<td>N129/748278: Inaha terrace</td>
<td>0.016</td>
<td>23</td>
</tr>
<tr>
<td>BJP-018X</td>
<td>N129/208944: Rapanui Terrace</td>
<td>0.034</td>
<td>93</td>
</tr>
<tr>
<td>BJP-020X</td>
<td>N129/927319: Brunswick Terrace</td>
<td>0.060</td>
<td>195</td>
</tr>
<tr>
<td>BJP-021B</td>
<td>N129/927319: Brunswick Terrace</td>
<td>0.041</td>
<td>120</td>
</tr>
<tr>
<td>BJP-022X</td>
<td>N129/927319: Brunswick Terrace</td>
<td>0.051</td>
<td>159</td>
</tr>
<tr>
<td>BJP-026</td>
<td>N129/013299: Kaitea Terrace (Meremere Lignite)</td>
<td>.118</td>
<td>400ka calibration</td>
</tr>
<tr>
<td>BJP-027</td>
<td>N129/013299: Kaitea Terrace (Meremere Lignite)</td>
<td>.111</td>
<td>.112 mean</td>
</tr>
<tr>
<td>BJP-028</td>
<td>N129/013299: Kaitea Terrace (Meremere Lignite)</td>
<td>.115</td>
<td>initial $D/L = 0.01$</td>
</tr>
<tr>
<td>BJP-029</td>
<td>N129/013299: Kaitea Terrace (Meremere Lignite)</td>
<td>.106</td>
<td>$K_L = 2.59 \times 10^{-7}$ yr$^{-1}$</td>
</tr>
<tr>
<td>BJP-030F</td>
<td>N129/793269: Rapanui Terrace</td>
<td>0.044</td>
<td>132</td>
</tr>
<tr>
<td>BJP-031</td>
<td>N129/793269: Rapanui Terrace</td>
<td>&gt;0.034</td>
<td>&gt;93</td>
</tr>
<tr>
<td>BJP-032</td>
<td>N129/793269: Rapanui Terrace</td>
<td>&gt;0.039</td>
<td>&gt;113</td>
</tr>
<tr>
<td>BJP-033</td>
<td>N129/877261: Rapanui Terrace</td>
<td>0.030</td>
<td>78</td>
</tr>
<tr>
<td>BJP-034B/034X</td>
<td>N129/910210: Rapanui Terrace</td>
<td>0.041/0.038</td>
<td>115(mean)</td>
</tr>
<tr>
<td>BJP-035</td>
<td>N129/910210: Rapanui Terrace</td>
<td>0.041</td>
<td>120</td>
</tr>
<tr>
<td>BJP-037</td>
<td>N137/132193: Post-Inaha Terrace</td>
<td>0.014</td>
<td>15</td>
</tr>
<tr>
<td>BJP-040</td>
<td>N137/159385: Post-Inaha Terrace</td>
<td>0.017</td>
<td>27</td>
</tr>
<tr>
<td>BJP-041</td>
<td>N137/18984: Post-Inaha Terrace (Marine)</td>
<td>0.026</td>
<td>62</td>
</tr>
<tr>
<td>BJP-042B</td>
<td>N137/350952: Rapanui Terrace (Rapanui Lignite)</td>
<td>0.034</td>
<td>93</td>
</tr>
<tr>
<td>BJP-044/044B</td>
<td>N130/32193: Brunswick Terrace</td>
<td>0.043/0.045</td>
<td>132</td>
</tr>
<tr>
<td>BJP-045</td>
<td>N130/32193: Brunswick Terrace</td>
<td>0.048</td>
<td>148</td>
</tr>
<tr>
<td>BJP-048B</td>
<td>N118/403473: Lahar 2m above HM</td>
<td>0.012</td>
<td>8</td>
</tr>
<tr>
<td>BJP-049</td>
<td>N137/265067: Ngarino Terrace</td>
<td>0.058</td>
<td>187</td>
</tr>
<tr>
<td>BJP-050</td>
<td>N137/442933: Ngarino Terrace</td>
<td>0.056</td>
<td>179</td>
</tr>
<tr>
<td>BJP-060</td>
<td>N129/757277: Inaha Terrace (Marine)</td>
<td>0.030</td>
<td>78</td>
</tr>
</tbody>
</table>

* D/L ratio (L-isoleucine/D-alloisoleucine) of amino acids presumed to be entirely within residual proteins after pretreatment with 6N HCl for 24 hours at room temperature.
# TABLE 3.5: WOOD IDENTIFICATIONS BY R.N. PATEL, BOTANY DIVISION, DSIR, CHRISTCHURCH, NEW ZEALAND

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Identification</th>
</tr>
</thead>
<tbody>
<tr>
<td>BJP-010</td>
<td>Dacrydium bidwillii or less likely, D. biforme</td>
</tr>
<tr>
<td>BJP-015</td>
<td></td>
</tr>
<tr>
<td>BJP-024</td>
<td></td>
</tr>
<tr>
<td>BJP-025</td>
<td></td>
</tr>
<tr>
<td>BJP-030</td>
<td></td>
</tr>
<tr>
<td>-BJP-033</td>
<td>Dacrydium bidwillii or Podocarpus spicatus</td>
</tr>
<tr>
<td>BJP-020</td>
<td></td>
</tr>
<tr>
<td>BJP-021</td>
<td></td>
</tr>
<tr>
<td>BJP-022</td>
<td>Leptospermum</td>
</tr>
<tr>
<td>BJP-023</td>
<td></td>
</tr>
<tr>
<td>BJP-027</td>
<td></td>
</tr>
<tr>
<td>BJP-029</td>
<td>Epacris or Dracophyllum</td>
</tr>
<tr>
<td>BJP-031</td>
<td>likely to be Dracophyllum</td>
</tr>
<tr>
<td>BJP-017</td>
<td></td>
</tr>
<tr>
<td>BJP-018</td>
<td></td>
</tr>
<tr>
<td>BJP-026</td>
<td></td>
</tr>
<tr>
<td>BJP-028</td>
<td></td>
</tr>
<tr>
<td>BJP-032</td>
<td>material too poor for identification</td>
</tr>
<tr>
<td>INTERNATIONAL UNIT</td>
<td>SERIES</td>
</tr>
<tr>
<td>--------------------</td>
<td>--------</td>
</tr>
<tr>
<td>Holocene</td>
<td>Recent</td>
</tr>
<tr>
<td></td>
<td>Hawera</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Quaternary</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower Quaternary</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Pliocene</td>
<td></td>
</tr>
</tbody>
</table>

* Only those units directly relevant to this thesis have been named.
<table>
<thead>
<tr>
<th></th>
<th>Aeolian Dune</th>
<th>Loess</th>
<th>Beach (littoral)</th>
<th>Shallow Marine</th>
<th>Fluvial</th>
<th>Estuarine</th>
<th>Volcanic Lahar</th>
<th>Tephra</th>
</tr>
</thead>
<tbody>
<tr>
<td>OTHER</td>
<td>Interbedded tephras.</td>
<td>Laterally continuous beds</td>
<td>Channel fill morphology</td>
<td>Laterally continuous beds</td>
<td>Associated with volcanic ring plains</td>
<td>Mantle palaeo-landforms</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>deformation structures</td>
<td>Mantle palaeo-landforms</td>
<td>Over erosional unconformity</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### TABLE 5.1: TERRACE DEFORMATION

<table>
<thead>
<tr>
<th></th>
<th>1 Vertical offset Moumahaki Fault (nearest 5m)</th>
<th>2 Vertical offset Ridge Rd Fault (nearest 5m)</th>
<th>3 Vertical offset Waitotara Fault (nearest 5m)</th>
<th>4 Vertical offset Nukumaru Fault (nearest 5m)</th>
<th>5 Shore parallel tilt Traverse 6-Traverse surface (degrees)</th>
<th>6 Shore normal tilt of wave cut surface (degrees)</th>
<th>7 Mean strandline height ratio relative to Kaiatea III</th>
</tr>
</thead>
<tbody>
<tr>
<td>Post-Inaha</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>.040</td>
<td>.37(.10)</td>
</tr>
<tr>
<td>Inaha</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>.052</td>
<td>.45(.18)</td>
</tr>
<tr>
<td>Rapanui</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>.080</td>
<td>.54(.27)</td>
</tr>
<tr>
<td>Ngarino</td>
<td>40</td>
<td>15</td>
<td>20</td>
<td>25</td>
<td>.103</td>
<td>.76(.49)</td>
<td>.74</td>
</tr>
<tr>
<td>Brunswick</td>
<td>40</td>
<td>30</td>
<td>15</td>
<td>25</td>
<td>-</td>
<td>.79(.52)</td>
<td>.795</td>
</tr>
<tr>
<td>Upper Brunswick</td>
<td>40</td>
<td>80</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>.155</td>
<td>.88(.61)</td>
</tr>
<tr>
<td>Kaiatea III</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>.189</td>
<td>.189</td>
<td>1</td>
</tr>
<tr>
<td>Kaiatea II</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>.229</td>
<td>1.05(.78)</td>
<td>1.13</td>
</tr>
<tr>
<td>Kaiatea I</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>1.30</td>
</tr>
<tr>
<td>Pre-Kaiatea</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>1.49</td>
</tr>
<tr>
<td>Pre-Kaiatea</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>50</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

* Initial wave cut surface dip estimated to be .27° (see p.126). Value in brackets is tilt corrected for initial dip.
<table>
<thead>
<tr>
<th>Traverse</th>
<th>Post-Inaha</th>
<th>Inaha</th>
<th>Rapanui</th>
<th>Ngarino</th>
<th>Brunswick</th>
<th>Upper Brunswick III</th>
<th>Kaiatea I</th>
<th>Kaiatea II</th>
<th>Pre-Kaiatea</th>
<th>Pre-Kaiatea</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-</td>
<td>-</td>
<td>48</td>
<td>90</td>
<td>145</td>
<td>-</td>
<td>190</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>-</td>
<td>24</td>
<td>50</td>
<td>100</td>
<td>160</td>
<td>-</td>
<td>212</td>
<td>232</td>
<td>284</td>
<td>-</td>
</tr>
<tr>
<td>3</td>
<td>-</td>
<td>28</td>
<td>56</td>
<td>112</td>
<td>174</td>
<td>-</td>
<td>230</td>
<td>256</td>
<td>292</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
<td>-</td>
<td>32</td>
<td>64</td>
<td>120</td>
<td>184</td>
<td>-</td>
<td>242</td>
<td>274</td>
<td>316</td>
<td>340+</td>
</tr>
<tr>
<td>5</td>
<td>-</td>
<td>36</td>
<td>68</td>
<td>126</td>
<td>192</td>
<td>-</td>
<td>256</td>
<td>290</td>
<td>332</td>
<td>-</td>
</tr>
<tr>
<td>6</td>
<td>16</td>
<td>40</td>
<td>70</td>
<td>128</td>
<td>194</td>
<td>214</td>
<td>264</td>
<td>348</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>7</td>
<td>16</td>
<td>40</td>
<td>70</td>
<td>126</td>
<td>194</td>
<td>214</td>
<td>264</td>
<td>348</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>8</td>
<td>16</td>
<td>40</td>
<td>68</td>
<td>120*</td>
<td>190</td>
<td>-</td>
<td>264</td>
<td>345</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>9</td>
<td>-</td>
<td>36</td>
<td>68</td>
<td>116*</td>
<td>-</td>
<td>-</td>
<td>258</td>
<td>340</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>10</td>
<td>-</td>
<td>30</td>
<td>60</td>
<td>104*</td>
<td>184</td>
<td>-</td>
<td>248</td>
<td>328</td>
<td>320</td>
<td>-</td>
</tr>
<tr>
<td>11</td>
<td>-</td>
<td>-</td>
<td>42</td>
<td>80*</td>
<td>-</td>
<td>168</td>
<td>212</td>
<td>280</td>
<td>312</td>
<td>360</td>
</tr>
<tr>
<td>12</td>
<td>-</td>
<td>-</td>
<td>42</td>
<td>76*</td>
<td>152</td>
<td>166</td>
<td>212</td>
<td>276</td>
<td>312</td>
<td>-</td>
</tr>
<tr>
<td>13</td>
<td>-</td>
<td>-</td>
<td>40</td>
<td>-</td>
<td>148</td>
<td>164</td>
<td>206</td>
<td>272</td>
<td>310</td>
<td>-</td>
</tr>
<tr>
<td>14</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>144</td>
<td>-</td>
<td>200</td>
<td>264</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

* Possible younger Ngarino strandline
TABLE 5.3: Terrace Strandline Height Ratios (relative to Kaiatea III strandline height)

<table>
<thead>
<tr>
<th>Traverse</th>
<th>Post-Inaha</th>
<th>Inaha</th>
<th>Rapanui</th>
<th>Ngarino</th>
<th>Brunswick</th>
<th>Upper Brunswick</th>
<th>Kaiatea III</th>
<th>Kaiatea II</th>
<th>Kaiatea I</th>
<th>Pre-Kaiatea</th>
<th>Pre-Kaiatea</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-</td>
<td>-</td>
<td>0.25</td>
<td>0.47</td>
<td>0.76</td>
<td>-</td>
<td>1</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>-</td>
<td>0.11</td>
<td>0.24</td>
<td>0.47</td>
<td>0.75</td>
<td>-</td>
<td>1</td>
<td>1.09</td>
<td>1.25</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>3</td>
<td>-</td>
<td>0.12</td>
<td>0.24</td>
<td>0.49</td>
<td>0.76</td>
<td>-</td>
<td>1</td>
<td>1.11</td>
<td>1.27</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
<td>-</td>
<td>0.13</td>
<td>0.25</td>
<td>0.50</td>
<td>0.76</td>
<td>-</td>
<td>1</td>
<td>1.13</td>
<td>1.31</td>
<td>1.40+</td>
<td>-</td>
</tr>
<tr>
<td>5</td>
<td>-</td>
<td>0.14</td>
<td>0.26</td>
<td>0.49</td>
<td>0.74</td>
<td>-</td>
<td>1</td>
<td>1.12</td>
<td>1.29</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>6</td>
<td>0.06</td>
<td>0.15</td>
<td>0.27</td>
<td>0.48</td>
<td>0.73</td>
<td>0.81</td>
<td>1</td>
<td>-</td>
<td>1.32</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>7</td>
<td>0.06</td>
<td>0.15</td>
<td>0.25</td>
<td>0.45*</td>
<td>0.72</td>
<td>-</td>
<td>1</td>
<td>-</td>
<td>1.31</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>8</td>
<td>-</td>
<td>0.14</td>
<td>0.26</td>
<td>0.41*</td>
<td>-</td>
<td>-</td>
<td>1</td>
<td>-</td>
<td>1.32</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>9</td>
<td>-</td>
<td>0.12</td>
<td>0.23</td>
<td>0.42*</td>
<td>0.74</td>
<td>-</td>
<td>1</td>
<td>1.15</td>
<td>1.32</td>
<td>1.51</td>
<td>-</td>
</tr>
<tr>
<td>10</td>
<td>-</td>
<td>-</td>
<td>0.20</td>
<td>0.38*</td>
<td>-</td>
<td>0.79</td>
<td>1</td>
<td>1.15</td>
<td>1.32</td>
<td>1.51</td>
<td>1.70</td>
</tr>
<tr>
<td>11</td>
<td>-</td>
<td>-</td>
<td>0.20</td>
<td>0.36*</td>
<td>0.72</td>
<td>0.78</td>
<td>1</td>
<td>1.13</td>
<td>1.30</td>
<td>1.47</td>
<td>1.70</td>
</tr>
<tr>
<td>12</td>
<td>-</td>
<td>-</td>
<td>0.19</td>
<td>-</td>
<td>0.72</td>
<td>0.80</td>
<td>1</td>
<td>1.15</td>
<td>1.32</td>
<td>1.50</td>
<td>-</td>
</tr>
<tr>
<td>13</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.72</td>
<td>-</td>
<td>1</td>
<td>1.15</td>
<td>1.32</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>14</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Mean: 0.06  0.13  0.24  0.45  0.74  0.795  1  1.13  1.30  1.49  1.70

* Possible younger Ngarino strandline
### TABLE 5.4: RACEMISATION DATES FROM RAPANUI TERRACE COVER BEDS

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Location</th>
<th>Bound D/L ratio</th>
<th>Age (Ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BJP-018</td>
<td>N129/944208 Manawapou River</td>
<td>0.034</td>
<td>93</td>
</tr>
<tr>
<td>BJP-030</td>
<td></td>
<td>0.044</td>
<td>132</td>
</tr>
<tr>
<td>BJP-031</td>
<td>N129/793269 Ohawe Beach</td>
<td>&gt;0.034</td>
<td>&gt;93</td>
</tr>
<tr>
<td>BJP-032</td>
<td></td>
<td>0.039</td>
<td>113</td>
</tr>
<tr>
<td>BJP-033</td>
<td>N129/877261 Whareroa</td>
<td>0.030</td>
<td>78</td>
</tr>
<tr>
<td>BJP-034</td>
<td>N129/900210 Tangahoe Beach</td>
<td>0.039</td>
<td>115</td>
</tr>
<tr>
<td>BJP-035</td>
<td>N129/900210 Tangahoe Beach</td>
<td>0.041</td>
<td>120</td>
</tr>
<tr>
<td>BJP-042</td>
<td>N137/350952 Ototoka West Beach</td>
<td>0.034</td>
<td>93</td>
</tr>
<tr>
<td>Mean age</td>
<td></td>
<td></td>
<td>105</td>
</tr>
<tr>
<td>Mean excluding BJP-033</td>
<td></td>
<td></td>
<td>110</td>
</tr>
<tr>
<td>BJP-001</td>
<td>N129/786289 Waingongoro River Terrace incised below Rapanui wave cut surface</td>
<td>0.029</td>
<td>74</td>
</tr>
<tr>
<td>Terrace</td>
<td>Racemisation/fission track age</td>
<td>Nukumaru Fault zone constant vertical offset age</td>
<td>Constant shore-parallel tilting age</td>
</tr>
<tr>
<td>---------------</td>
<td>--------------------------------</td>
<td>-----------------------------------------------</td>
<td>---------------------------------</td>
</tr>
<tr>
<td>Post-Inaha</td>
<td>60</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Inaha</td>
<td>80-100</td>
<td>-</td>
<td>100</td>
</tr>
<tr>
<td>Rapanui</td>
<td>&gt;110 (120)</td>
<td>200±70</td>
<td>130</td>
</tr>
<tr>
<td>Ngarino</td>
<td>&gt;185 (200)</td>
<td>270±90</td>
<td>210</td>
</tr>
<tr>
<td>Brunswick</td>
<td>&gt;200 &lt;340</td>
<td>330±110</td>
<td>270</td>
</tr>
<tr>
<td>Upper Brunswick</td>
<td>&lt;370</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Kaiatea III</td>
<td>400</td>
<td>400 (ref)</td>
<td>400 (ref)</td>
</tr>
<tr>
<td>Kaiatea II</td>
<td>&gt;400</td>
<td>400±130</td>
<td>490</td>
</tr>
<tr>
<td>Kaiatea I</td>
<td>&gt;400</td>
<td>530±180</td>
<td>590</td>
</tr>
<tr>
<td>Pre-Kaiatea</td>
<td>&gt;400</td>
<td>670±220</td>
<td>-</td>
</tr>
<tr>
<td>Pre-Kaiatea</td>
<td>&gt;400</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

* Based on uniform shore-normal uplift rates; will under-estimate ages of younger terraces because tilt has occurred.
### TABLE 5.6: OVERSEAS SEA LEVEL CHRONOLOGY: Ages of relative high sea level events (Ka); paleo-sea-level estimates relative to present given where known (metres).

<table>
<thead>
<tr>
<th>Barbados*</th>
<th>Huon Peninsula, New Guinea+</th>
<th>Atauro and Timor#</th>
<th>Honshu Japan**</th>
<th>Ryukyu Islands##</th>
<th>California++</th>
</tr>
</thead>
<tbody>
<tr>
<td>30Ka(-40m)</td>
<td>40-50Ka(-30m)</td>
<td></td>
<td>42Ka</td>
<td></td>
<td></td>
</tr>
<tr>
<td>60Ka</td>
<td>60Ka(-28m)</td>
<td>80Ka(+6m)</td>
<td>67Ka</td>
<td>80Ka(-10m)</td>
<td></td>
</tr>
<tr>
<td>80Ka(-15m)</td>
<td>80Ka(-15m)</td>
<td>100Ka(-13m)</td>
<td>80Ka(-3 to -9m)</td>
<td>100Ka(-3 to -9m)</td>
<td>90Ka(-10m)</td>
</tr>
<tr>
<td>100Ka(-13m)</td>
<td></td>
<td></td>
<td></td>
<td>100Ka(-3 to -9m)</td>
<td>100Ka(-10m)</td>
</tr>
<tr>
<td>125Ka(+5m)</td>
<td>120Ka</td>
<td>120Ka(+5 to +8m)</td>
<td>130Ka(+5m)</td>
<td>120Ka</td>
<td></td>
</tr>
<tr>
<td>135Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>180Ka</td>
<td>185Ka(-25m)</td>
<td></td>
<td>175Ka(-20m)</td>
<td>190Ka(-4m)</td>
<td>200Ka</td>
</tr>
<tr>
<td>200Ka</td>
<td>220Ka</td>
<td>220Ka(-20m to -30m)</td>
<td>190Ka(-4m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>220Ka</td>
<td>240Ka</td>
<td></td>
<td>220-230Ka(-10m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>250Ka</td>
<td>280Ka</td>
<td></td>
<td>270-280Ka(-10m)</td>
<td></td>
<td>300Ka</td>
</tr>
<tr>
<td>300Ka</td>
<td>320Ka</td>
<td>340Ka</td>
<td>350Ka(3m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>320Ka</td>
<td>360Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>350Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>380Ka</td>
<td>420Ka</td>
<td>440Ka</td>
<td>400Ka</td>
<td></td>
<td></td>
</tr>
<tr>
<td>460Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>490Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>520Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>590Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>640Ka</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* James et al (1971), Matthews (1973), Bender et al (1979)
# Chappell & Veeh (1978a)
** Machida (1973)
## Konishi et al (1970)
TABLE 5.7: UPLIFT RATES (mm/yr), HEIGHTS (m) AND ESTIMATED AGES (Ka) OF THE NGARINO, BRUNSWICK AND UPPER BRUNSWICK STRANDLINES (Hinging-type uplift model)

<table>
<thead>
<tr>
<th>Traverse No.</th>
<th>Ngarino</th>
<th>Brunswick</th>
<th>Upper Brunswick</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.41</td>
<td>0.45</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>0.43</td>
<td>0.49</td>
<td>-</td>
</tr>
<tr>
<td>3</td>
<td>0.48</td>
<td>0.55</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
<td>0.54</td>
<td>0.60</td>
<td>-</td>
</tr>
<tr>
<td>5</td>
<td>0.58</td>
<td>0.62</td>
<td>0.64</td>
</tr>
<tr>
<td>6</td>
<td>0.61</td>
<td>0.63</td>
<td>0.64</td>
</tr>
<tr>
<td>7</td>
<td>0.60</td>
<td>0.63</td>
<td>0.64</td>
</tr>
<tr>
<td>8</td>
<td>0.60</td>
<td>0.64</td>
<td>-</td>
</tr>
<tr>
<td>9</td>
<td>0.60</td>
<td>0.64</td>
<td>-</td>
</tr>
<tr>
<td>10</td>
<td>0.52</td>
<td>0.59</td>
<td>0.59</td>
</tr>
<tr>
<td>11</td>
<td>0.38</td>
<td>0.47</td>
<td>0.48</td>
</tr>
<tr>
<td>12</td>
<td>0.38</td>
<td>0.48</td>
<td>0.49</td>
</tr>
<tr>
<td>13</td>
<td>0.35</td>
<td>0.49</td>
<td>0.50</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Traverse No.</th>
<th>Ngarino</th>
<th>Brunswick</th>
<th>Upper Brunswick</th>
</tr>
</thead>
<tbody>
<tr>
<td>14</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Mean age c.210 Ka BP
Mean age c.310 Ka BP
Mean age c.340 Ka BP

# mean age c.220 Ka BP
* mean age c.200 Ka BP
TABLE 5.8: UPLIFT RATES (mm/year) STRANDLINE HEIGHTS (m) AND PALEOSEALEVEL ESTIMATES (m) FOR THE INAHA AND POST-INAHA AND POST INAHA TERRACES ASSUMING CORRELATION WITH c.105Ka AND c.84Ka BP INTERSTADIAL HIGH SEA LEVELS

Model (A): Uniform uplift, equal to uplift rate on Rapanui strandline
Model (B): Hinging uplift, extrapolated from Rapanui and Kaiatea III strandlines

<table>
<thead>
<tr>
<th>Traverse No.</th>
<th>Inaha strandline 105Ka BP?</th>
<th>Post-Inaha strandline 84Ka BP?</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Strandline Height (A)</td>
<td>Uplift Rate (A)</td>
</tr>
<tr>
<td>1</td>
<td>24</td>
<td>.38</td>
</tr>
<tr>
<td>2</td>
<td>28</td>
<td>.43</td>
</tr>
<tr>
<td>3</td>
<td>32</td>
<td>.49</td>
</tr>
<tr>
<td>4</td>
<td>36</td>
<td>.53</td>
</tr>
<tr>
<td>5</td>
<td>40</td>
<td>.54</td>
</tr>
<tr>
<td>6</td>
<td>40</td>
<td>.54</td>
</tr>
<tr>
<td>7</td>
<td>40</td>
<td>.53</td>
</tr>
<tr>
<td>8</td>
<td>36</td>
<td>.53</td>
</tr>
<tr>
<td>9</td>
<td>30</td>
<td>.46</td>
</tr>
<tr>
<td>10</td>
<td>30</td>
<td>.46</td>
</tr>
<tr>
<td>11</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>12/13</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>14</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Mean</td>
<td>-18m</td>
<td>-13m</td>
</tr>
</tbody>
</table>
TABLE 5.9: UPLIFT RATES (mm/year), STRANDLINE HEIGHTS (m) AND AGE ESTIMATES (Ka) FOR THE KAIATEA II, KAIATEA I AND PRE-KAIATEA TERRACES (paleosealevel equal to present)

Model (A): Uniform uplift, equal to uplift rate on Kaiatea III strandline.
Model (B): Hinging uplift, extrapolated from Rapanui and Kaiatea III strandlines.

<table>
<thead>
<tr>
<th>Traverse No.</th>
<th>Kaiatea II strandline</th>
<th>Kaiatea I strandline</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Strandline Height</td>
<td>Uplift Rate (A)</td>
</tr>
<tr>
<td>1</td>
<td>190</td>
<td>.48</td>
</tr>
<tr>
<td>2</td>
<td>212</td>
<td>.53</td>
</tr>
<tr>
<td>3</td>
<td>230</td>
<td>.58</td>
</tr>
<tr>
<td>4</td>
<td>242</td>
<td>.61</td>
</tr>
<tr>
<td>5</td>
<td>258</td>
<td>.65</td>
</tr>
<tr>
<td>6</td>
<td>264</td>
<td>.66</td>
</tr>
<tr>
<td>7</td>
<td>264</td>
<td>.66</td>
</tr>
<tr>
<td>8</td>
<td>258</td>
<td>.65</td>
</tr>
<tr>
<td>9</td>
<td>248</td>
<td>.62</td>
</tr>
<tr>
<td>10</td>
<td>212</td>
<td>.53</td>
</tr>
<tr>
<td>11</td>
<td>212</td>
<td>.53</td>
</tr>
<tr>
<td>12</td>
<td>206</td>
<td>.52</td>
</tr>
<tr>
<td>13</td>
<td>200</td>
<td>.50</td>
</tr>
<tr>
<td>Mean ages</td>
<td></td>
<td>c.450</td>
</tr>
<tr>
<td>Younger pre-Kaiatea strandline</td>
<td></td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>320</td>
<td>.53</td>
</tr>
<tr>
<td>12</td>
<td>312</td>
<td>.53</td>
</tr>
<tr>
<td>13</td>
<td>310</td>
<td>.52</td>
</tr>
<tr>
<td>Mean ages</td>
<td></td>
<td>c.595</td>
</tr>
<tr>
<td>Terrace</td>
<td>Strandline Age (Ka)</td>
<td>Basis</td>
</tr>
<tr>
<td>------------</td>
<td>---------------------</td>
<td>--------------------------------------------</td>
</tr>
<tr>
<td>Post-Inaha</td>
<td>60</td>
<td>Hinging uplift model</td>
</tr>
<tr>
<td>Post-Inaha</td>
<td>80</td>
<td>Racemisation/hinging uplift model</td>
</tr>
<tr>
<td>Inaha</td>
<td>100</td>
<td>Racemisation/hinging uplift model</td>
</tr>
<tr>
<td>Rapanui</td>
<td>120</td>
<td>Racemisation/overseas data</td>
</tr>
<tr>
<td>Ngarino</td>
<td>210*</td>
<td>Racemisation/hinging uplift model</td>
</tr>
<tr>
<td>Brunswick</td>
<td>310</td>
<td>Hinging uplift model</td>
</tr>
<tr>
<td>Upper Brunswick</td>
<td>340</td>
<td>Hinging uplift model</td>
</tr>
<tr>
<td>Kaiatea III</td>
<td>400</td>
<td>Relation to Omahina Tephra (370±50Ka BP)</td>
</tr>
<tr>
<td>Kaiatea II</td>
<td>440</td>
<td>Uniform uplift model</td>
</tr>
<tr>
<td>Kaiatea I</td>
<td>520</td>
<td>Uniform uplift model</td>
</tr>
<tr>
<td>Pre-Kaiatea</td>
<td>595</td>
<td>Uniform uplift model</td>
</tr>
<tr>
<td>Pre-Kaiatea</td>
<td>680</td>
<td>Uniform uplift model</td>
</tr>
</tbody>
</table>

* Possibly composite feature; upper strandline 220Ka BP, lower strandline 200Ka BP
<table>
<thead>
<tr>
<th></th>
<th>Wanganui Valley (Fleming 1953, 1957)</th>
<th>Rangitikei Valley* (Te Punga 1952, Fleming 1957, Milne 1973a, b)</th>
<th>South Taranaki Marine terraces (this study)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>HAHRA SERIES</strong></td>
<td>Mt Curl Tephra-----------------------</td>
<td>Ngarino Terrace</td>
<td></td>
</tr>
<tr>
<td><strong>Landguard Formation</strong></td>
<td>Mingaroa Fossil Beds</td>
<td>Rangitawa Pumice -----------------------</td>
<td>Omahina Tephra 370±40Ka</td>
</tr>
<tr>
<td><strong>Putiki Shellbed</strong></td>
<td>Mosstown Sand</td>
<td>Putiki Shellbeds</td>
<td>Kakariki Conglomerate</td>
</tr>
<tr>
<td><strong>Karaka Siltstone</strong></td>
<td>Upper Castlecliff Shellbed</td>
<td>Ruamahanga Conglomerate</td>
<td></td>
</tr>
<tr>
<td><strong>Shakespeare Cliff Siltstone</strong></td>
<td>Shakespeare Cliff Sand</td>
<td>Tainui Shellbed</td>
<td>Onepuhi Shellbeds-Kaiatea III Terrace</td>
</tr>
<tr>
<td><strong>Pinnacle Sand</strong></td>
<td>Lower Castlecliff Shellbed</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Seafield Sand</strong></td>
<td>Upper Kai-Iwi Siltstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Kai-Iwi GROUP</strong></td>
<td>Kupe Formation</td>
<td>Waioimo Shellbed 450±90Ka</td>
<td>Kaiatea II Terrace</td>
</tr>
<tr>
<td></td>
<td>Upper Westmere Siltstone</td>
<td>Waitapu Shellbed 520±80Ka</td>
<td>Kaiatea I Terrace</td>
</tr>
<tr>
<td></td>
<td>Omapu Shellbed</td>
<td>Kaukatea Ash 570±80Ka</td>
<td>Pre-Kaiatea Terraces</td>
</tr>
<tr>
<td></td>
<td>Lower Kai-Iwi Siltstone</td>
<td>Pataka Pumice 610±60; 640±180Ka</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Kaimatira Pumice Sand</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Brunhes/Matuyama paleomagnetic boundary</strong></td>
<td>Rewa Pumice 740±90Ka</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Fission track dates from Seaward (1974, 1976, 1979), Boellstorff & Te Punga (1977)
<table>
<thead>
<tr>
<th>Lahar and source (Grant-Taylor 1964a, b; Neall 1979)</th>
<th>South Taranaki marine terrace and age estimate</th>
<th>Input of black sand to South Taranaki marine terraces</th>
<th>Volcanic activity in western Taranaki</th>
<th>K/Ar ages, western Taranaki lavas (Stipp 1968; Neall 1979)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Opunake (E)</td>
<td>60Ka unnamed 80Ka unnamed Inaha 105Ka</td>
<td>high</td>
<td>EGMONT</td>
<td>21±3Ka</td>
</tr>
<tr>
<td>Stratford (E)</td>
<td>Rapanui 120Ka</td>
<td>high</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lepperton (E and/or P)</td>
<td>Ngarino 210Ka</td>
<td>high</td>
<td>POUAKAI</td>
<td>224±7Ka 216±7Ka 227±7Ka 246±4Ka 249±3Ka</td>
</tr>
<tr>
<td>Maitahi (P)</td>
<td>Brunswick 310Ka Upper Brunswick 340Ka</td>
<td>high</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inglewood (K)*</td>
<td>Kaiatea III 400Ka</td>
<td>low</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eltham (K)*</td>
<td>Kaiatea II 440Ka</td>
<td>low</td>
<td></td>
<td></td>
</tr>
<tr>
<td>New Plymouth (K)*</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inglewood? (K) Eltham (K)# New Plymouth? (K)</td>
<td>Kaiatea I 520Ka 2 pre Kaiatea terraces 600Ka/680Ka</td>
<td>low-moderate</td>
<td>KAITAKE</td>
<td>575±14Ka</td>
</tr>
</tbody>
</table>

* postulated stratigraphic position, Grant-Taylor 1964a, b.
# approximate stratigraphic position, this study.
TABLE 5.13: PRESENT NEW ZEALAND UPPER QUATERNARY CHRONOLOGY BASED ON CLIMATIC SUBDIVISION (after Suggate 1978)

<table>
<thead>
<tr>
<th>STAGE of WARMING and WARMTH</th>
<th>STAGE of COOLING and COLD</th>
<th>CULMINATING CLIMATIC EVENT</th>
<th>ORIGIN OF NAME</th>
<th>FIRST DEFINITION</th>
<th>LATEST DEFINITION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Araniuian</td>
<td>present-day warmth</td>
<td>Aranui (Christchurch); area of thick sediment consequent upon rising post-glacial sea level</td>
<td>Suggate 1962</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Otiran</td>
<td>Otira Glacial</td>
<td>Otira, on main divide of Southern Alps</td>
<td>Gage &amp; Suggate 1958</td>
<td>Suggate 1965</td>
<td></td>
</tr>
<tr>
<td>Oturian</td>
<td>Oturi Interglacial</td>
<td>Oturi Lake, near Waverley; warm-water marine fauna</td>
<td>Fleming 1953</td>
<td>Suggate 1965</td>
<td></td>
</tr>
<tr>
<td>Waiamean</td>
<td>Waiamea Glacial</td>
<td>Waiamea, north Westland, in region of type glacial deposits</td>
<td>Suggate 1965</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Terangian</td>
<td>Terangi Interglacial</td>
<td>Terangi; Maori name at Mt Jowett. Wanganui; warm-water marine fauna</td>
<td>Fleming 1953</td>
<td>Suggate 1965</td>
<td></td>
</tr>
<tr>
<td>Waimaungan</td>
<td>Waimaunga Glacial</td>
<td>Waimaunga, southwest Nelson, near type glacial deposits</td>
<td>Gage &amp; Suggate 1958</td>
<td>Suggate 1965</td>
<td></td>
</tr>
<tr>
<td>Waiwheran</td>
<td>Waiwhero Interglacial</td>
<td>Waiwhero Creek, north Westland, at type interglacial beach deposits</td>
<td>Suggate 1965</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Porikan</td>
<td>Porika Glacial</td>
<td>Porika track, upper Buller valley, which crosses type glacial deposits</td>
<td>Gage, 1961</td>
<td>Suggate 1965</td>
<td></td>
</tr>
<tr>
<td>South Taranaki Terraces</td>
<td>Present Stages</td>
<td>Modified Stages A</td>
<td>Modified Stages B</td>
<td></td>
<td></td>
</tr>
<tr>
<td>-------------------------</td>
<td>----------------</td>
<td>-------------------</td>
<td>-------------------</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Aranui</td>
<td>Aranui</td>
<td>Aranui</td>
<td></td>
<td></td>
</tr>
<tr>
<td>60Ka BP unnamed</td>
<td></td>
<td>Otiran</td>
<td>Otiran</td>
<td></td>
<td></td>
</tr>
<tr>
<td>80Ka BP unnamed</td>
<td>Otiran</td>
<td>Oturian</td>
<td>Oturian</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inaha (Stratford Lahars)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rapanui (Lepperton Lahars)</td>
<td></td>
<td>Oturian</td>
<td>Waimean</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ngarino (Maitahi Lahars)</td>
<td></td>
<td>Waimean</td>
<td>Terangian</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Brunswick Upper Brunswick</td>
<td></td>
<td>Terangian</td>
<td>Waiheran</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Waimaungan</td>
<td>Waimaungan</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kaiatea III</td>
<td>Waiheran</td>
<td>Waiheran</td>
<td>Castlecliffian</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kaiatea II Kaiatea I 2 Pre-Kaiatea terraces</td>
<td>Castlecliffian?</td>
<td>Castlecliffian</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

(A) Oturian Stage tied to type section
Terangian Stage tied to type section

(B) Oturian Stage correlated with last interglacial (sensu lato)
Terangian Stage correlated with penultimate interglacial
<table>
<thead>
<tr>
<th>VALLEY NUMBER</th>
<th>VALLEY AGE (km)</th>
<th>TOTAL THAMES ACCRETE AREA, ( d^2 ) (km²)</th>
<th>TOTAL STORM AREA, ( d^2 ) (km²)</th>
<th>TOTAL VALLEY AREA, ( d^2 ) (km²)</th>
<th>ALTITUDE, ( m )</th>
<th>LENGTH, ( m )</th>
<th>WIDTH, ( m )</th>
<th>RELIEF RATIO</th>
<th>SLOPE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>300</td>
<td>45436</td>
<td>5732</td>
<td>179440</td>
<td>1230</td>
<td>1140</td>
<td>86.5</td>
<td>70</td>
<td>14.5</td>
</tr>
<tr>
<td>4</td>
<td>200</td>
<td>15272</td>
<td>8160</td>
<td>48864</td>
<td>512</td>
<td>416</td>
<td>49</td>
<td>30</td>
<td>19</td>
</tr>
<tr>
<td>5</td>
<td>200</td>
<td>18600</td>
<td>9496</td>
<td>59296</td>
<td>656</td>
<td>566</td>
<td>51</td>
<td>24</td>
<td>19</td>
</tr>
<tr>
<td>6</td>
<td>200</td>
<td>38376</td>
<td>17888</td>
<td>122960</td>
<td>1238</td>
<td>1152</td>
<td>63</td>
<td>53</td>
<td>10</td>
</tr>
<tr>
<td>7</td>
<td>200</td>
<td>29520</td>
<td>15504</td>
<td>96112</td>
<td>858</td>
<td>798</td>
<td>57</td>
<td>44.5</td>
<td>13*</td>
</tr>
<tr>
<td>8</td>
<td>300</td>
<td>33576</td>
<td>7672</td>
<td>146608</td>
<td>834</td>
<td>754</td>
<td>83</td>
<td>70</td>
<td>13</td>
</tr>
<tr>
<td>9</td>
<td>300</td>
<td>90424</td>
<td>5696</td>
<td>120368</td>
<td>656</td>
<td>566</td>
<td>94</td>
<td>66.5</td>
<td>27.5</td>
</tr>
<tr>
<td>10</td>
<td>200</td>
<td>50160</td>
<td>33640</td>
<td>196416</td>
<td>1092</td>
<td>1018</td>
<td>79</td>
<td>57</td>
<td>22</td>
</tr>
<tr>
<td>11</td>
<td>200</td>
<td>8936</td>
<td>7148</td>
<td>42320</td>
<td>354</td>
<td>203</td>
<td>59</td>
<td>19</td>
<td>46</td>
</tr>
<tr>
<td>12</td>
<td>200</td>
<td>5424</td>
<td>3482</td>
<td>26708</td>
<td>228</td>
<td>82</td>
<td>46</td>
<td>8</td>
<td>38</td>
</tr>
<tr>
<td>13</td>
<td>200</td>
<td>9640</td>
<td>3758</td>
<td>37600</td>
<td>324</td>
<td>308</td>
<td>54</td>
<td>42</td>
<td>12</td>
</tr>
<tr>
<td>14</td>
<td>120</td>
<td>5120</td>
<td>2132</td>
<td>15916</td>
<td>226</td>
<td>190</td>
<td>49</td>
<td>37</td>
<td>12</td>
</tr>
<tr>
<td>15</td>
<td>15732</td>
<td>8400</td>
<td>62648</td>
<td>586</td>
<td>502</td>
<td>50</td>
<td>35</td>
<td>35*</td>
<td>15</td>
</tr>
<tr>
<td>16</td>
<td>12410</td>
<td>6476</td>
<td>45912</td>
<td>426</td>
<td>408</td>
<td>47.5</td>
<td>37</td>
<td>10.5</td>
<td>18</td>
</tr>
<tr>
<td>17</td>
<td>6212</td>
<td>2034</td>
<td>34232</td>
<td>371</td>
<td>233</td>
<td>12</td>
<td>21</td>
<td>158</td>
<td>143</td>
</tr>
<tr>
<td>18</td>
<td>5878</td>
<td>1560</td>
<td>26368</td>
<td>285</td>
<td>269</td>
<td>40</td>
<td>29.5</td>
<td>10.5</td>
<td>16</td>
</tr>
<tr>
<td>19</td>
<td>15092</td>
<td>7922</td>
<td>64688</td>
<td>583</td>
<td>505</td>
<td>59</td>
<td>49</td>
<td>10*</td>
<td>39*</td>
</tr>
<tr>
<td>20</td>
<td>7330</td>
<td>3192</td>
<td>19625</td>
<td>444</td>
<td>333</td>
<td>41</td>
<td>29</td>
<td>12*</td>
<td>37*</td>
</tr>
<tr>
<td>21</td>
<td>787</td>
<td>257</td>
<td>4350</td>
<td>85</td>
<td>45</td>
<td>22</td>
<td>8.5</td>
<td>13.5</td>
<td>40</td>
</tr>
<tr>
<td>22</td>
<td>5248</td>
<td>1418</td>
<td>15787</td>
<td>289</td>
<td>232</td>
<td>36</td>
<td>24</td>
<td>12*</td>
<td>28*</td>
</tr>
<tr>
<td>23</td>
<td>1026</td>
<td>57</td>
<td>5321</td>
<td>79</td>
<td>59</td>
<td>33</td>
<td>17</td>
<td>16</td>
<td>20</td>
</tr>
<tr>
<td>24</td>
<td>300</td>
<td>9539</td>
<td>1782</td>
<td>22307</td>
<td>220</td>
<td>178</td>
<td>42</td>
<td>18.5</td>
<td>23.5</td>
</tr>
<tr>
<td>25</td>
<td>200</td>
<td>2914</td>
<td>863</td>
<td>14095</td>
<td>161</td>
<td>140</td>
<td>28</td>
<td>15</td>
<td>13</td>
</tr>
<tr>
<td>26</td>
<td>10708</td>
<td>52</td>
<td>43912</td>
<td>476</td>
<td>431</td>
<td>43</td>
<td>27.5</td>
<td>15.5</td>
<td>45</td>
</tr>
<tr>
<td>27</td>
<td>5682</td>
<td>0</td>
<td>2404</td>
<td>516</td>
<td>488</td>
<td>21</td>
<td>15</td>
<td>6</td>
<td>28</td>
</tr>
</tbody>
</table>

| MEAN          | 216           | 15403                                      | 6173                          | 59050             | 509     | 440    | 50     | 33.5      | 16.9  |

* Mean value
|       | $V_{th}$ | $t$   | $V_m$ | $A$   | $L_v$ | $L_s$ | $F_v$ | $F_s$ | $H$   | $D_o$ | $N$   | $RR$  | $S$   | $W$   | $D$   | $DD$  | $r_d$ | $r_h$ | $C$   | $CH$  | $M$   | $V_{th}/L_v^2$ |
|-------|---------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|----------------|
| $t$   | .36     |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $V_m$ | .79     | -.05  |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $A$   | .99     | .41   |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $L_v$ | .95     | .23   | .73   |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $L_s$ | .94     | .23   | .72   |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $F_v$ | .85     | .58   | .56   | .87   | .73   |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $F_s$ | .85     | .48   | .52   | .86   | .78   |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $H$   | .04     | .26   | .10   | .09   | .08   | -.17  | .28   |       |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $D_o$ | .24     | .15   | .20   | .28   | .24   | .12   | .33   | -.01  |       |       |       |       |       |       |       |       |       |       |       |       |                 |
| $N$   | -.07    | -.16  | .03   | -.11  | .04   | .02   | -.08  | .04   | -.26  | -.28  |       |       |       |       |       |       |       |       |       |       |                 |
| $RR$  | .52     | .02   | -.42  | -.49  | -.66  | -.66  | -.32  | -.40  | .19   | -.18  | -.23  |       |       |       |       |       |       |       |       |       |                 |
| $S$   | -.45    | .02   | -.40  | -.43  | -.61  | -.57  | -.24  | -.22  | -.03  | -.33  | -.13  | .95   |       |       |       |       |       |       |       |       |                 |
| $W$   | .37     | .25   | .51   | .41   | .20   | .18   | .52   | .39   | .34   | .24   | -.20  | .11   | -.13  |       |       |       |       |       |       |       |                 |
| $D$   | .85     | .51   | .68   | .86   | .72   | .71   | .93   | .85   | .24   | .26   | -.14  | -.40  | -.35  | .66   |       |       |       |       |       |       |                 |
| $DD$  | -.59    | -.52  | -.47  | -.64  | -.45  | -.44  | -.74  | -.57  | -.42  | .39   | .40   | .18   | .23   | -.68  | .81   |       |       |       |       |       |                 |
| $r_d$ | .75     | -.05  | -.22  | -.72  | .69   | .68   | .68   | .66   | .12   | .18   | -.06  | .61   | -.45  | .60   | .82   |   .62  |       |       |       |       |                 |
| $r_h$ | .70     | -.16  | .63   | .68   | .80   | .81   | .42   | .55   | -.25  | .04   | .14   | -.60  | -.53  | .13   | .40   | -.16  | .57   |       |       |       |                 |
| $C$   | .61     | .79   | .09   | .65   | .52   | .54   | .75   | .78   | .02   | .03   | -.22  | -.23  | -.14  | .31   | .79   | .54   | .28   | .22  |       |                 |
| $CH$  | -.25    | -.17  | -.05  | -.18  | -.37  | -.39  | -.06  | -.13  | .20   | .01   | .11   | .41   | .44   | -.02  | -.15  | -.01  | -.08  | -.20  | -.34  |       |                 |
| $M$   | -.14    | -.27  | .55   | .11   | .11   | .05   | .18   | -.02  | .43   | .38   | .18   | -.16  | -.21  | .37   | .27   | -.30  | .52   | .05   | -.95  | .33   |                 |
| $V_{th}/L_v^2$ | -.58 | .06  | -.43  | -.53  | -.74  | -.71  | -.35  | -.45  | .22   | .21   | -.31  | .95   | .86   | -.03  | .35   | .09   | -.46  | .70   | -.22  | .38   | -.10   |
| $U$   | -.41    | .19   | -.22  | -.35  | -.52  | -.52  | -.23  | -.28  | .15   | -.14  | .16   | .47   | .44   | -.02  | -.25  | -.00  | -.38  | -.46  | -.30  | .76   | .15   | .50   |

---

*Significant at 99.9% confidence level*
TABLE 6.4: Log-log multiple regression, normalised thalweg slice area (V<sub>th/L_v</sub><sup>2</sup>)

<table>
<thead>
<tr>
<th>log V&lt;sub&gt;th/L_v&lt;/sub&gt;&lt;sup&gt;2&lt;/sup&gt;</th>
<th>log t</th>
<th>log M</th>
<th>log C</th>
<th>log CH</th>
<th>log RR</th>
<th>log F&lt;sub&gt;v&lt;/sub&gt;</th>
<th>log L_v</th>
<th>log A</th>
</tr>
</thead>
<tbody>
<tr>
<td>log V&lt;sub&gt;th/L_v&lt;/sub&gt;&lt;sup&gt;2&lt;/sup&gt;</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>log t</td>
<td>.22</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>log M</td>
<td>-.64</td>
<td>.43</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>log C</td>
<td>-.83</td>
<td>.60</td>
<td>-.08</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>log CH</td>
<td>-.25</td>
<td>.15</td>
<td>.52</td>
<td>-.28</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>log RR</td>
<td>.96</td>
<td>.23</td>
<td>.31</td>
<td>-.27</td>
<td>.55</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>log F&lt;sub&gt;v&lt;/sub&gt;</td>
<td>-.22</td>
<td>.59</td>
<td>.43</td>
<td>.60</td>
<td>.09</td>
<td>-.22</td>
<td></td>
<td></td>
</tr>
<tr>
<td>log L_v</td>
<td>.46</td>
<td>.15</td>
<td>-.00</td>
<td>.52</td>
<td>-.36</td>
<td>-.85</td>
<td>.69</td>
<td>.94</td>
</tr>
<tr>
<td>log A</td>
<td>.34</td>
<td>.33</td>
<td>.14</td>
<td>.60</td>
<td>.23</td>
<td>-.67</td>
<td>.80</td>
<td>.94</td>
</tr>
</tbody>
</table>

(B) Regression equations

1. Valley age (t) and valley length (L_v)
   \[ V_{th/L_v} = 0.04t^{0.79}L_v^{-0.63} \quad R^2 = 0.818 \]
   \[ p(Ho) \text{ total } <0.005; \quad t <0.005; \quad L_v <0.005 \]

2. Valley age (t) and valley length (L_v) and mudstone ratio (M)
   \[ V_{th/L_v} = 0.14t^{0.55}L_v^{0.62}F_v^{0.05} \quad R^2 = 0.864 \]
   \[ p(Ho) \text{ total } <0.005; \quad t <0.025; \quad L_v <0.005; \quad M <0.025 \]

3. Relief Ratio (RR)
   \[ V_{th/L_v} = 0.43R^{0.92} \quad R^2 = 0.924 \]
   \[ p(Ho) <0.005 \]

4. All variables \[ R^2 = 0.974 \quad p(Ho) \text{ total regression } <0.005 \]
<table>
<thead>
<tr>
<th>Site</th>
<th>R²</th>
<th>a</th>
<th>b</th>
</tr>
</thead>
<tbody>
<tr>
<td>SV027</td>
<td>0.98</td>
<td>1.48</td>
<td>0.426</td>
</tr>
<tr>
<td>SV026</td>
<td>0.97</td>
<td>3.21</td>
<td>0.412</td>
</tr>
<tr>
<td>SV025</td>
<td>0.98</td>
<td>4.55</td>
<td>0.361</td>
</tr>
<tr>
<td>SV024</td>
<td>1.00</td>
<td>6.27</td>
<td>0.355</td>
</tr>
<tr>
<td>SV023</td>
<td>0.98</td>
<td>3.99</td>
<td>0.469</td>
</tr>
<tr>
<td>SV022</td>
<td>0.96</td>
<td>0.78</td>
<td>0.719</td>
</tr>
<tr>
<td>SV021</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>SV020</td>
<td>0.94</td>
<td>1.47</td>
<td>0.666</td>
</tr>
<tr>
<td>SV019</td>
<td>1.00</td>
<td>0.54</td>
<td>0.751</td>
</tr>
<tr>
<td>SV018</td>
<td>0.99</td>
<td>3.09</td>
<td>0.458</td>
</tr>
<tr>
<td>SV017</td>
<td>1.00</td>
<td>1.59</td>
<td>0.513</td>
</tr>
<tr>
<td>SV016</td>
<td>0.98</td>
<td>2.94</td>
<td>0.462</td>
</tr>
<tr>
<td>SV015</td>
<td>0.95</td>
<td>3.56</td>
<td>0.429</td>
</tr>
<tr>
<td>SV014</td>
<td>0.99</td>
<td>4.11</td>
<td>0.455</td>
</tr>
<tr>
<td>SV013</td>
<td>0.99</td>
<td>2.85</td>
<td>0.504</td>
</tr>
<tr>
<td>SV012</td>
<td>0.96</td>
<td>8.56</td>
<td>0.316</td>
</tr>
<tr>
<td>SV011</td>
<td>0.97</td>
<td>4.38</td>
<td>0.448</td>
</tr>
<tr>
<td>SV010</td>
<td>0.98</td>
<td>2.56</td>
<td>0.494</td>
</tr>
<tr>
<td>SV009</td>
<td>0.98</td>
<td>1.49</td>
<td>0.637</td>
</tr>
<tr>
<td>SV008</td>
<td>0.95</td>
<td>1.30</td>
<td>0.608</td>
</tr>
<tr>
<td>SV007</td>
<td>0.99</td>
<td>2.64</td>
<td>0.465</td>
</tr>
<tr>
<td>SV006</td>
<td>0.98</td>
<td>0.40</td>
<td>0.717</td>
</tr>
<tr>
<td>SV005</td>
<td>0.97</td>
<td>1.63</td>
<td>0.529</td>
</tr>
<tr>
<td>SV004</td>
<td>0.94</td>
<td>1.07</td>
<td>0.625</td>
</tr>
<tr>
<td>SV003A</td>
<td>0.93</td>
<td>0.33</td>
<td>0.795</td>
</tr>
<tr>
<td>SV003</td>
<td>0.96</td>
<td>0.00</td>
<td>1.400</td>
</tr>
<tr>
<td>SV002A</td>
<td>0.98</td>
<td>0.08</td>
<td>1.083</td>
</tr>
<tr>
<td>SV002</td>
<td>0.98</td>
<td>0.19</td>
<td>0.886</td>
</tr>
<tr>
<td>SV001</td>
<td>0.92</td>
<td>0.88</td>
<td>0.623</td>
</tr>
</tbody>
</table>

\[ Y = ax^b \]
TABLE 6.6: Slope data, summary

<table>
<thead>
<tr>
<th></th>
<th>Mean Gradient</th>
<th>Standard Deviation</th>
<th>Mean Vertical Height</th>
<th>Mean Horizontal Length</th>
<th>Mean Ground Surface Length</th>
<th>Kurtosis</th>
<th>Skewness</th>
<th>No. of Observations</th>
<th>No. of Profiles</th>
</tr>
</thead>
<tbody>
<tr>
<td>All Slopes</td>
<td>28.2°</td>
<td>9.1°</td>
<td>25.4m</td>
<td>47.4m</td>
<td>54.6m</td>
<td>4.66</td>
<td>0.50</td>
<td>5060</td>
<td>139</td>
</tr>
<tr>
<td>200Ka</td>
<td>28.5°</td>
<td>8.7°</td>
<td>24.3m</td>
<td>45.0m</td>
<td>51.9m</td>
<td>4.74</td>
<td>0.59</td>
<td>3288</td>
<td>95</td>
</tr>
<tr>
<td>300Ka</td>
<td>28.0°</td>
<td>9.8°</td>
<td>27.8m</td>
<td>52.4m</td>
<td>60.4m</td>
<td>4.44</td>
<td>0.38</td>
<td>1772</td>
<td>44</td>
</tr>
<tr>
<td>West facing (east side)</td>
<td>27.7°</td>
<td>9.0°</td>
<td>25.1m</td>
<td>47.8m</td>
<td>54.9m</td>
<td>5.32</td>
<td>0.61</td>
<td>2413</td>
<td>66</td>
</tr>
<tr>
<td>East facing (west side)</td>
<td>28.8°</td>
<td>9.3°</td>
<td>25.4m</td>
<td>46.1m</td>
<td>53.5m</td>
<td>4.19</td>
<td>0.48</td>
<td>2390</td>
<td>67</td>
</tr>
<tr>
<td>Cover beds segment</td>
<td>26.8°</td>
<td>11.8°</td>
<td>16.0m</td>
<td>15.4m</td>
<td>26.2m</td>
<td>3.73</td>
<td>1.43</td>
<td>3357</td>
<td>139</td>
</tr>
<tr>
<td>Mudstone segment</td>
<td>31.4°</td>
<td>17.2°</td>
<td>9.4m</td>
<td>32.0m</td>
<td>18.4m</td>
<td>3.21</td>
<td>1.56</td>
<td>1703</td>
<td>139</td>
</tr>
</tbody>
</table>
TABLE 6.7: Mann-Whitney U-test of slope asymmetry for paired profiles (see Appendix 4 for locations)

<table>
<thead>
<tr>
<th>Profile No.</th>
<th>No significant (p(H₀)&gt;0.05)</th>
<th>W-facing steeper (p(H₀)&lt;0.5)</th>
<th>E-facing steeper (p(H₀)&lt;0.5)</th>
</tr>
</thead>
<tbody>
<tr>
<td>22</td>
<td>71</td>
<td>49</td>
<td>25</td>
</tr>
<tr>
<td>23</td>
<td>72</td>
<td>49</td>
<td>31</td>
</tr>
<tr>
<td>24#</td>
<td>73</td>
<td>76</td>
<td>40#</td>
</tr>
<tr>
<td>28</td>
<td>78</td>
<td>87#</td>
<td>41</td>
</tr>
<tr>
<td>30</td>
<td>82</td>
<td>93</td>
<td>44</td>
</tr>
<tr>
<td>35</td>
<td>86</td>
<td>99*</td>
<td>46</td>
</tr>
<tr>
<td>38</td>
<td>88</td>
<td></td>
<td>47</td>
</tr>
<tr>
<td>39</td>
<td>91</td>
<td></td>
<td>51</td>
</tr>
<tr>
<td>42</td>
<td>92</td>
<td></td>
<td>53/54</td>
</tr>
<tr>
<td>45</td>
<td>94</td>
<td></td>
<td>55</td>
</tr>
<tr>
<td>48</td>
<td>96</td>
<td></td>
<td>58</td>
</tr>
<tr>
<td>50</td>
<td>97</td>
<td></td>
<td>79</td>
</tr>
<tr>
<td>56</td>
<td>98</td>
<td></td>
<td>83*</td>
</tr>
<tr>
<td>59</td>
<td></td>
<td></td>
<td>89</td>
</tr>
<tr>
<td>62</td>
<td></td>
<td>South facing steeper</td>
<td></td>
</tr>
<tr>
<td>63</td>
<td></td>
<td>North facing steeper</td>
<td>32</td>
</tr>
<tr>
<td>64/65</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>66</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* East facing slope expected to be steeper
# West facing slope expected to be steeper

orientation significantly different from E-W for wave cut surfaces dips to be influencing slope morphology.
APPENDIX 1 - Amino Acid Analysis

Locations of samples (otherwise as for Table 3.4).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>BJP-005</td>
<td>N129/918195 Fluvial terrace, mouth of Manawapou River</td>
</tr>
<tr>
<td>BJP-007</td>
<td>N129/773276 Lahar, Ohawe Beach</td>
</tr>
<tr>
<td>BJP-011</td>
<td>N129/748278 Inaha type section</td>
</tr>
<tr>
<td>BJP-012P</td>
<td>N129/748278 Inaha type section</td>
</tr>
<tr>
<td>BJP-016P</td>
<td>N129/748278 Inaha type section</td>
</tr>
<tr>
<td>BJP-017</td>
<td>N129/208944 Rapanui Terrace cover beds</td>
</tr>
<tr>
<td>BJP-019</td>
<td>N129/208944 Rapanui Terrace cover beds</td>
</tr>
<tr>
<td>BJP-038</td>
<td>N137/185985 Tree stump in growth position on beach face</td>
</tr>
<tr>
<td>BJP-039</td>
<td>N137/185985 Channel deposit at HWM</td>
</tr>
<tr>
<td>BJP-043</td>
<td>N137/393937 Butlers Shell Conglomerate (peat)</td>
</tr>
<tr>
<td>BJP-053</td>
<td>Rotorua area Underlying (?) Rotoehu Ash (location unknown)</td>
</tr>
</tbody>
</table>

**KEY**

LYS - Lysine; HIS - Hystidine; ARG - Arginine; ASP - Aspartic acid;  
THR - Threonine; SER - Serine; GLU - Glutamic acid; PRO - Proline;  
GLY - Glycine; ALA - Alanine; CYS - Cysteine; VAL - Valine;  
MET - Methionine; ILEU - Isoleucine; LEV - Leucine; TYR - Tyrosine;  
PHE - Phenylalanine; ALEU - alloisoleucine

U - small unknown peak eluting between buffer and Methionine  
A - small unknown peak eluting between Methionine and alloisoleucine  
D/L - ratio of D-alloisoleucine/L-isoleucine  
µM/g - amino acid concentration in micromoles/gram of wood or peat  
OH-Pro - Hydroxyproline; ORN - Ornithine; ß-ala - ß-alanine

* present but not measured  
? identification uncertain

Concentrations of individual amino acids measured as a percentage of total ASP + THR + SER + GLU + PRO + GLY + ALA + VAL + MET + ILEU + LEU
<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
<th>14</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>001X</td>
<td>002X</td>
<td>004X</td>
<td>004X</td>
<td>005X</td>
<td>006X</td>
<td>007X</td>
<td>010X</td>
<td>011X</td>
<td>012P</td>
<td>012PX</td>
<td>014X</td>
<td>015X</td>
<td></td>
</tr>
<tr>
<td></td>
<td>BOUND</td>
<td>BOUND</td>
<td>FREE</td>
<td>BOUND</td>
<td>FREE</td>
<td>BOUND</td>
<td>FREE</td>
<td>BOUND</td>
<td>BOUND</td>
<td>FREE</td>
<td>BOUND</td>
<td>FREE</td>
<td>BOUND</td>
<td></td>
</tr>
<tr>
<td>LYS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>HIS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>ARG</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>ASP</td>
<td>11.8</td>
<td>10.1</td>
<td>8.4</td>
<td>10.4</td>
<td>10.9</td>
<td>10.4</td>
<td>11.5</td>
<td>9.5</td>
<td>11.3</td>
<td>7.6</td>
<td>16.2</td>
<td>13.9</td>
<td>15.6</td>
<td>11.3</td>
</tr>
<tr>
<td>THR</td>
<td>7.3</td>
<td>7.0</td>
<td>8.6</td>
<td>7.7</td>
<td>11.6</td>
<td>9.5</td>
<td>7.5</td>
<td>8.3</td>
<td>7.1</td>
<td>6.7</td>
<td>8.3</td>
<td>10.2</td>
<td>7.3</td>
<td>6.5</td>
</tr>
<tr>
<td>SER</td>
<td>9.2</td>
<td>10.1</td>
<td>9.4</td>
<td>8.4</td>
<td>13.8</td>
<td>11.9</td>
<td>10.5</td>
<td>13.1</td>
<td>8.8</td>
<td>9.0</td>
<td>7.3</td>
<td>6.4</td>
<td>7.6</td>
<td>10.8</td>
</tr>
<tr>
<td>GLU</td>
<td>10.5</td>
<td>11.6</td>
<td>12.4</td>
<td>10.9</td>
<td>10.9</td>
<td>13.0</td>
<td>11.8</td>
<td>10.8</td>
<td>7.8</td>
<td>8.5</td>
<td>13.1</td>
<td>10.3</td>
<td>10.7</td>
<td>9.1</td>
</tr>
<tr>
<td>PRO</td>
<td>7.8</td>
<td>8.2</td>
<td>4.9</td>
<td>11.1</td>
<td>6.2</td>
<td>6.8</td>
<td>10.9</td>
<td>7.5</td>
<td>8.9</td>
<td>8.3</td>
<td>3.6</td>
<td>6.5</td>
<td>5.8</td>
<td>15.9</td>
</tr>
<tr>
<td>GLY</td>
<td>18.8</td>
<td>19.0</td>
<td>14.8</td>
<td>14.5</td>
<td>15.8</td>
<td>16.7</td>
<td>16.3</td>
<td>16.9</td>
<td>17.1</td>
<td>20.9</td>
<td>24.1</td>
<td>20.6</td>
<td>19.7</td>
<td>15.7</td>
</tr>
<tr>
<td>ALA</td>
<td>16.2</td>
<td>16.8</td>
<td>29.4</td>
<td>18.3</td>
<td>11.1</td>
<td>13.8</td>
<td>14.6</td>
<td>12.4</td>
<td>17.8</td>
<td>19.6</td>
<td>15.8</td>
<td>16.9</td>
<td>16.8</td>
<td>13.1</td>
</tr>
<tr>
<td>CYS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>VAL</td>
<td>4.9</td>
<td>4.8</td>
<td>4.2</td>
<td>4.3</td>
<td>6.6</td>
<td>6.2</td>
<td>4.9</td>
<td>5.9</td>
<td>6.0</td>
<td>5.6</td>
<td>4.4</td>
<td>5.6</td>
<td>4.8</td>
<td>5.6</td>
</tr>
<tr>
<td>MET</td>
<td>1.7</td>
<td>1.2</td>
<td>0.7</td>
<td>1.2</td>
<td>1.4</td>
<td>1.1</td>
<td>1.3</td>
<td>1.5</td>
<td>2.2</td>
<td>1.7</td>
<td>2.2</td>
<td>1.4</td>
<td>3.7</td>
<td>1.7</td>
</tr>
<tr>
<td>ILEU</td>
<td>2.8</td>
<td>2.9</td>
<td>2.7</td>
<td>3.2</td>
<td>4.2</td>
<td>3.7</td>
<td>2.6</td>
<td>3.2</td>
<td>3.0</td>
<td>2.6</td>
<td>2.3</td>
<td>2.2</td>
<td>2.0</td>
<td>2.6</td>
</tr>
<tr>
<td>LEU</td>
<td>9.1</td>
<td>8.3</td>
<td>4.5</td>
<td>8.5</td>
<td>7.4</td>
<td>6.9</td>
<td>8.1</td>
<td>6.8</td>
<td>9.9</td>
<td>9.6</td>
<td>2.7</td>
<td>6.1</td>
<td>5.9</td>
<td>7.9</td>
</tr>
<tr>
<td>TYR</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>?</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>PHE</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>?</td>
<td>?</td>
<td>*</td>
<td>*</td>
<td>?</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>ALEU</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>trace</td>
<td>*</td>
<td>trace</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>U</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>*</td>
</tr>
<tr>
<td>A</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>*</td>
</tr>
<tr>
<td>D/L</td>
<td>.029</td>
<td>nd</td>
<td>.083</td>
<td>&gt;.012</td>
<td>nd</td>
<td>.034</td>
<td>nd</td>
<td>&gt;.022</td>
<td>nd</td>
<td>.018</td>
<td>nd</td>
<td>.016</td>
<td></td>
<td></td>
</tr>
<tr>
<td>wM/g</td>
<td>23.7</td>
<td>23.1</td>
<td>1.3</td>
<td>24.3</td>
<td>2.7</td>
<td>1.7</td>
<td>25.4</td>
<td>.7</td>
<td>32.9</td>
<td>11.5</td>
<td>16.0</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>16</td>
<td>17</td>
<td>18</td>
<td>19</td>
<td>20</td>
<td>21</td>
<td>22</td>
<td>23</td>
<td>24</td>
<td>25</td>
<td>26</td>
<td>27</td>
<td>28</td>
<td></td>
</tr>
<tr>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td>----</td>
<td></td>
</tr>
<tr>
<td>016P FREE</td>
<td>16PX FREE</td>
<td>017 FREE</td>
<td>018X FREE</td>
<td>009X FREE</td>
<td>020 FREE</td>
<td>020X FREE</td>
<td>021 FREE</td>
<td>021X FREE</td>
<td>022X FREE</td>
<td>022X FREE</td>
<td>026 FREE</td>
<td>026X FREE</td>
<td>027X FREE</td>
<td>028 FREE</td>
</tr>
<tr>
<td>LYS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>HIS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>ARG</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>ASP</td>
<td>9.4</td>
<td>11.7</td>
<td>12.7</td>
<td>10.0</td>
<td>10.6</td>
<td>6.4</td>
<td>10.7</td>
<td>6.6</td>
<td>9.7</td>
<td>10.0</td>
<td>7.9</td>
<td>9.6</td>
<td>7.4</td>
<td>7.5</td>
</tr>
<tr>
<td>THR</td>
<td>7.3</td>
<td>10.0</td>
<td>9.8</td>
<td>8.5</td>
<td>8.2</td>
<td>7.0</td>
<td>8.1</td>
<td>5.2</td>
<td>7.9</td>
<td>5.2</td>
<td>6.3</td>
<td>5.3</td>
<td>4.0</td>
<td>5.3</td>
</tr>
<tr>
<td>SER</td>
<td>6.7</td>
<td>10.0</td>
<td>10.1</td>
<td>15.9</td>
<td>5.8</td>
<td>10.7</td>
<td>11.7</td>
<td>6.4</td>
<td>8.5</td>
<td>7.2</td>
<td>10.0</td>
<td>8.0</td>
<td>9.3</td>
<td>7.7</td>
</tr>
<tr>
<td>GLU</td>
<td>7.9</td>
<td>13.6</td>
<td>12.0</td>
<td>8.8</td>
<td>10.5</td>
<td>9.0</td>
<td>10.8</td>
<td>7.8</td>
<td>10.7</td>
<td>9.0</td>
<td>9.7</td>
<td>7.7</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>PRO</td>
<td>7.2</td>
<td>12.7</td>
<td>7.1</td>
<td>9.9</td>
<td>11.8</td>
<td>7.4</td>
<td>9.5</td>
<td>6.6</td>
<td>7.4</td>
<td>9.3</td>
<td>14.5</td>
<td>7.7</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>GLY</td>
<td>3.0</td>
<td>15.9</td>
<td>16.0</td>
<td>15.1</td>
<td>15.1</td>
<td>16.4</td>
<td>16.1</td>
<td>14.3</td>
<td>17.3</td>
<td>16.0</td>
<td>11.2</td>
<td>17.5</td>
<td>14.2</td>
<td>14.4</td>
</tr>
<tr>
<td>ALA</td>
<td>7.6</td>
<td>11.9</td>
<td>13.4</td>
<td>15.5</td>
<td>15.4</td>
<td>16.5</td>
<td>15.5</td>
<td>17.6</td>
<td>16.0</td>
<td>24.6</td>
<td>12.5</td>
<td>19.6</td>
<td>13.9</td>
<td>16.3</td>
</tr>
<tr>
<td>CYS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>VAL</td>
<td>4.1</td>
<td>3.7</td>
<td>6.7</td>
<td>5.0</td>
<td>7.8</td>
<td>7.4</td>
<td>4.9</td>
<td>10.1</td>
<td>6.8</td>
<td>5.3</td>
<td>10.3</td>
<td>7.0</td>
<td>8.4</td>
<td>9.9</td>
</tr>
<tr>
<td>MET</td>
<td>2.7</td>
<td>2.6</td>
<td>1.5</td>
<td>2.1</td>
<td>1.7</td>
<td>2.5</td>
<td>1.7</td>
<td>2.8</td>
<td>2.0</td>
<td>1.4</td>
<td>1.6</td>
<td>0.7</td>
<td>2.9</td>
<td>2.5</td>
</tr>
<tr>
<td>ILEU</td>
<td>1.9</td>
<td>2.3</td>
<td>5.2</td>
<td>2.9</td>
<td>4.3</td>
<td>5.7</td>
<td>2.8</td>
<td>8.4</td>
<td>4.7</td>
<td>3.1</td>
<td>6.1</td>
<td>4.2</td>
<td>7.6</td>
<td>6.1</td>
</tr>
<tr>
<td>LEU</td>
<td>2.3</td>
<td>5.6</td>
<td>5.6</td>
<td>6.3</td>
<td>8.8</td>
<td>11.1</td>
<td>8.1</td>
<td>14.1</td>
<td>8.8</td>
<td>10.6</td>
<td>9.9</td>
<td>12.7</td>
<td>9.6</td>
<td>12.4</td>
</tr>
<tr>
<td>TYR</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>PHE</td>
<td>?</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>ALEU</td>
<td>trace</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>A</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>trace</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>D/L</td>
<td>nd</td>
<td>nd</td>
<td>.26</td>
<td>.034</td>
<td>nd</td>
<td>.033</td>
<td>.060</td>
<td>.107</td>
<td>.051</td>
<td>.118</td>
<td>.111</td>
<td>.115</td>
<td>.106</td>
<td>-</td>
</tr>
<tr>
<td>μM/g</td>
<td>0.8</td>
<td>60.7</td>
<td>1.5</td>
<td>15.6</td>
<td>16.3</td>
<td>1.1</td>
<td>8.8</td>
<td>1.2</td>
<td>8.3</td>
<td>2.9</td>
<td>4.0</td>
<td>2.1</td>
<td>1.1</td>
<td>1.2</td>
</tr>
<tr>
<td></td>
<td>29</td>
<td>30</td>
<td>31</td>
<td>32</td>
<td>33</td>
<td>34</td>
<td>35</td>
<td>36</td>
<td>37</td>
<td>38</td>
<td>39</td>
<td>40</td>
<td>41</td>
<td>42</td>
</tr>
<tr>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td></td>
<td>030</td>
<td>030AP</td>
<td>030B</td>
<td>030CP</td>
<td>030D</td>
<td>030F</td>
<td>031X</td>
<td>031</td>
<td>032</td>
<td>033</td>
<td>034B</td>
<td>034X</td>
<td>035B</td>
<td>037</td>
</tr>
<tr>
<td>LYS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>HIS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>ARG</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>ASP</td>
<td>7.6</td>
<td>8.4</td>
<td>16.0</td>
<td>9.8</td>
<td>9.8</td>
<td>12.0</td>
<td>7.5</td>
<td>10.8</td>
<td>8.7</td>
<td>9.5</td>
<td>10.3</td>
<td>9.6</td>
<td>11.1</td>
<td>9.4</td>
</tr>
<tr>
<td>THR</td>
<td>6.7</td>
<td>7.7</td>
<td>12.8</td>
<td>10.4</td>
<td>13.7</td>
<td>8.6</td>
<td>7.0</td>
<td>4.8</td>
<td>6.8</td>
<td>7.3</td>
<td>8.0</td>
<td>5.7</td>
<td>6.6</td>
<td>7.5</td>
</tr>
<tr>
<td>SER</td>
<td>6.4</td>
<td>7.3</td>
<td>13.6</td>
<td>8.6</td>
<td>12.6</td>
<td>8.2</td>
<td>6.5</td>
<td>8.2</td>
<td>8.8</td>
<td>8.0</td>
<td>8.4</td>
<td>11.4</td>
<td>13.9</td>
<td>6.2</td>
</tr>
<tr>
<td>GLU</td>
<td>8.2</td>
<td>23.9</td>
<td>22.0</td>
<td>10.9</td>
<td>11.3</td>
<td>9.2</td>
<td>8.7</td>
<td>6.8</td>
<td>9.0</td>
<td>10.9</td>
<td>11.4</td>
<td>9.9</td>
<td>7.4</td>
<td>9.2</td>
</tr>
<tr>
<td>PRO</td>
<td>10.0</td>
<td>9.1</td>
<td>trace</td>
<td>-</td>
<td>-</td>
<td>6.0</td>
<td>10.3</td>
<td>18.1</td>
<td>20.5</td>
<td>15.2</td>
<td>9.0</td>
<td>12.7</td>
<td>12.1</td>
<td>13.5</td>
</tr>
<tr>
<td>GLY</td>
<td>13.4</td>
<td>17.6</td>
<td>15.0</td>
<td>21.2</td>
<td>19.1</td>
<td>18.1</td>
<td>13.5</td>
<td>12.2</td>
<td>11.2</td>
<td>13.2</td>
<td>14.9</td>
<td>12.3</td>
<td>14.6</td>
<td>11.3</td>
</tr>
<tr>
<td>ALA</td>
<td>17.6</td>
<td>13.6</td>
<td>12.5</td>
<td>22.9</td>
<td>15.6</td>
<td>15.9</td>
<td>15.2</td>
<td>11.1</td>
<td>10.5</td>
<td>12.3</td>
<td>14.3</td>
<td>12.5</td>
<td>16.5</td>
<td>14.3</td>
</tr>
<tr>
<td>CYS</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>VAL</td>
<td>10.3</td>
<td>4.1</td>
<td>4.3</td>
<td>8.4</td>
<td>7.1</td>
<td>8.3</td>
<td>11.5</td>
<td>5.4</td>
<td>8.4</td>
<td>8.6</td>
<td>8.3</td>
<td>8.5</td>
<td>5.8</td>
<td>11.9</td>
</tr>
<tr>
<td>MET</td>
<td>1.8</td>
<td>3.8</td>
<td>trace</td>
<td>1.2</td>
<td>0.9</td>
<td>1.4</td>
<td>1.8</td>
<td>1.2</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
<td>0.9</td>
<td>1.7</td>
</tr>
<tr>
<td>ILEU</td>
<td>6.3</td>
<td>2.2</td>
<td>1.4</td>
<td>2.6</td>
<td>3.8</td>
<td>5.1</td>
<td>6.5</td>
<td>3.3</td>
<td>5.1</td>
<td>4.7</td>
<td>6.4</td>
<td>5.5</td>
<td>3.8</td>
<td>6.5</td>
</tr>
<tr>
<td>LEU</td>
<td>11.6</td>
<td>2.5</td>
<td>2.2</td>
<td>3.9</td>
<td>6.2</td>
<td>7.3</td>
<td>11.5</td>
<td>8.3</td>
<td>8.9</td>
<td>8.3</td>
<td>12.1</td>
<td>10.0</td>
<td>9.4</td>
<td>10.4</td>
</tr>
<tr>
<td>TYR</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>PHE</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>ALLEU</td>
<td>*</td>
<td>trace</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>U</td>
<td>*</td>
<td>*</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>A</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>trace</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>D/L</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>µM/g</td>
<td>4.2</td>
<td>0.3</td>
<td>0.06</td>
<td>2.5</td>
<td>0.6</td>
<td>49.1</td>
<td>13.0</td>
<td>5.1</td>
<td>5.9</td>
<td>4.0</td>
<td>9.4</td>
<td>10.1</td>
<td>2.1</td>
<td>12.7</td>
</tr>
<tr>
<td>OH-Pro</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>ORN</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Am-Sugar</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>β-alal</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

# 24 hours in 6NHC1 at room temperature
+ 85 hours in 6NHC1 at room temperature
<table>
<thead>
<tr>
<th>44</th>
<th>45</th>
<th>46</th>
<th>47</th>
<th>48</th>
<th>49</th>
<th>50</th>
<th>51</th>
<th>52</th>
<th>53</th>
<th>54</th>
<th>55</th>
<th>56</th>
<th>57</th>
<th>58</th>
</tr>
</thead>
<tbody>
<tr>
<td>43</td>
<td>039</td>
<td>040</td>
<td>041</td>
<td>042</td>
<td>043</td>
<td>044</td>
<td>045</td>
<td>046</td>
<td>047</td>
<td>048</td>
<td>049</td>
<td>050</td>
<td>051</td>
<td>052</td>
</tr>
<tr>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
<td><strong>BOUND</strong></td>
</tr>
<tr>
<td>LYS</td>
<td>ARG</td>
<td>ASN</td>
<td>ASP</td>
<td>GRF</td>
<td>GLU</td>
<td>GLY</td>
<td>PRO</td>
<td>GLY</td>
<td>ALA</td>
<td>VAL</td>
<td>ILE</td>
<td>LEU</td>
<td>PHE</td>
<td>ALEU</td>
</tr>
<tr>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>12.0</td>
<td>12.5</td>
<td>12.6</td>
<td>12.8</td>
<td>13.1</td>
<td>13.3</td>
<td>13.5</td>
<td>13.7</td>
<td>14.0</td>
<td>14.2</td>
<td>14.5</td>
<td>14.8</td>
<td>15.0</td>
<td>15.2</td>
<td>15.5</td>
</tr>
</tbody>
</table>

**Notes:**
- * indicates a significant increase.
- ? indicates data not available.

**Reference Values:**
- LYS: 10.0
- ARG: 12.0
- ASN: 14.0
- ASP: 16.0
- GRF: 18.0
- GLU: 20.0
- GLY: 22.0
- PRO: 24.0
- GLY: 26.0
- ALA: 28.0
- VAL: 30.0
- ILE: 32.0
- LEU: 34.0
- PHE: 36.0
- ALEU: 38.0

**Additional Information:**
- D/L-Glu: 2.0
- OH-Pro: 4.0
- Am-Sugar: 6.0
- Sarc: 8.0
- OH-Lys: 10.0
- A-O-Lys: 12.0

**Calculation:**
- Normal range = 10.0 - 12.0
- High range = 12.1 - 14.0
- Very high range = 14.1 - 16.0
APPENDIX 2 - Localities where Omahina Tephra has been identified.

1. N109/979939: Kaipakari Rd; white tephra poorly exposed within weathered dark brown ashes; Elevation unknown.
2. N119/904423: Rotokare Rd; 15cm white clay with sandy basal layer; beneath chocolate-brown andesitic ash; elevation 198m; Kaiatea III marine terrace.
3. N129/948367: Morea Rd; 15cm white clay with sandy basal layer; elevation 210m; Kaiatea III marine terrace (see Section A).
4. N129/979339: Farmtrack off Tangahoe Valley Rd; thin white ash; elevation 218m; Kaiatea III marine terrace.
5. N129/999320: Farm tracks north of Meremere Rd; thin white ash; elevation unknown; Kaiatea III marine terrace.
6. N129/013299: Soil slip, eastern side of tributary to Otoki Stream; 15cm white ash; elevation 257m; Kaiatea II marine terrace (see Section B).
7. N129/031259: Farm track north of Ingahape Rd; thin white ash beneath chocolate-brown ash; elevation 218m; Kaiatea III marine terrace.
8. N129/087238: Ball Rd Extension; thin white ash within dark-brown ashes; elevation unknown; Kaiatea I marine terrace.
9. N130/122178: Elliot Rd; prominent white ash within weathered brown ashes; elevation 256m; Kaiatea III marine terrace.
10. N130/184160: Farmtrack and soil slips east of Kohi Rd; 15cm white ash with sandy basal layer; beneath black dune-sand; elevation 243m; Kaiatea III marine terrace.
11. N130/237150: Omahina Rd; 15cm white tephra with sandy basal layer; fission-track sample locality; elevation 257m; Kaiatea III marine terrace (see Section C).
12. N130/296120: Soil slip north of landing strip, Ridge Rd; white ash beneath dark-brown weathered ash; elevation 203m; Kaiatea III marine terrace (see Section D).
13. N137/471066: Rangitatau East Rd south of school; thin white ash poorly exposed beneath red-brown and chocolate-brown weathered ash; elevation 359m; pre-Kaiatea I marine terrace.
14. N137/466044: Rangitatau East Rd; very thin white ash underlying prominent dark-brown ash; elevation 288m; Kaiatea I marine terrace.
15. N137/469035: Rangitatau East Rd; tephric sand beneath grey dune-sand; elevation 266m; Kaiatea I marine terrace.
16. N138/762873: Farm track north of Denlair Rd; 40cm weathered white ash; beneath grey dune-sand; elevation 208m; Kaiatea III (?) marine terrace.

Note: All heights reported have been measured to the nearest metre using a Paulin surveying altimeter. Estimated error less than 3m in all cases.
THE STRATIGRAPHIC POSITION OF THE MT CURL TEPHRA

At its type section on Mt Curl Road (N138/956822) the Mt Curl Tephra is a c.90cm thick sub-horizontal rhyolitic tephra layer which overlies loess and underlies dunesands (Milne 1973a). A fission track age of $230^{\pm}30$ Ka BP was reported by Milne (1973a) and has subsequently been confirmed by Pillans & Kohn (in press) (Table 4.3). The tephra on Mt Curl Road was first described by Fleming (1953) who correlated it with the Fordell Ash. However, Milne (1973a) concluded that for the following reasons, the Mt Curl Tephra and Fordell Ash were different beds:

1. Samples of supposed Fordell Ash at its type exposures (Fleming 1953, Figs. 49 and 50) consisted of alluvial silts containing "about 80% aggregates, probably clay, about 10% diatom skeletons, about 10% of quartz and less than 0.05% of volcanic glass. No ferromagnesian minerals were recognised" (Milne 1973a, p.523). That is, it appeared to be a diatomaceous earth rather than a tephra.

2. Although the stratigraphic positions of both Mt Curl Tephra and Fordell Ash were superficially quite similar (both directly overlain by dunesand), the sequence of loess beds in overlying sediments was apparently less complete at the Fordell ash type exposures, and quite different from the sequence at the Mt Curl type section. It was suggested by Milne (1973a, p.523) that the dunesands at the type exposures of the Fordell Ash were deposited during the cutting of the Ngarino marine terrace while those at Mt Curl type section were deposited during the cutting of the Brunswick Terrace.
Milne (1973a) argued that because of similarities in the loess sequence overlying the Mt Curl Tephra and that overlying the Brunswick Terrace (locality not specified), and because of the proximity of the Brunswick Marine terrace to the Mt Curl type section, it was reasonably certain that the dunesands overlying the Mt Curl Tephra were derived from the coastline during the time the Brunswick Terrace was forming i.e. the Terangi Interglacial. Therefore (Milne 1973a, p.527), since the Mt Curl Tephra (only slightly weathered) rested on cold climate deposits (loess) supposedly predating the Terangi Interglacial, and underlay dunesands deposited during the Terangi Interglacial, it was argued that the tephra fell near its beginning. The beginning of the Terangi Interglacial was therefore assigned on age of c.230Ka BP, while its end was placed at c.190Ka BP (based on estimated weathering rates on dunesand at the Mt Curl Type Section and a fission track age of 230±30Ka BP for the Mt Curl Tephra). In other words the Brunswick wave cut surface was supposedly formed between 190 and 230Ka BP. Height relationships, reported by Milne (1973a), between the Brunswick and Ngarino Terraces, suggested that the Ngarino Terrace stopped forming at or more recently than 115Ka BP.

The above age estimates are at variance with those proposed in this thesis for the Brunswick and Ngarino Terraces, which are here assigned ages of c.310 and 210Ka respectively (see Table 5.10). However, Milne's work can be questioned on several counts as follows:

(1) Milne (1973a, p.525) stated that "there is no evidence to suggest that any of the older coastal dunesands in the area were deposited during the last glaciation and it has been therefore assumed that all coastal dune sands are likely to be of interglacial origin" (In support of his correlation of the dunesands overlying the Mt Curl Tephra with the cutting of the Brunswick Terrace during the Terangi Interglacial). However, as indicated earlier (p. 36) the Aokoutere Ash (Kawakawa Tephra Formation) which is dated at circa 20,000 BP is interbedded with dunesands at some localities, and clearly indicates that dune formation occurred during the last glaciation.
(2) He correlated the dunesands overlying the Mt Curl Tephra with Brunswick Dunesand (following Fleming (1953)), which at its type section overlies, and is clearly younger than, the Brunswick marine terrace. Since both Fleming (1953) and Milne (1973a) have accepted that dunesands overlying the Brunswick Terrace probably relate to the cutting of the Ngarino Terrace, the logical conclusion is that either

(a) the dunesands overlying the Mt Curl Tephra relate to the cutting of the Ngarino Terrace, or

(b) the dunesands overlying the Mt Curl Tephra are not Brunswick Dunesand.

(3) My own observations at the type exposures of the Fordell Ash indicate the presence of at least 2 prominent rhyolitic tephras, either of which might correlate with the Mt Curl (work is currently in progress to clarify this problem).

(4) Milne (1973a, p.523) notes that the cover bed sequence of loess beds at the Fordell Ash type exposures has been eroded off the dunesands, but that "a full sequence is clearly exposed in a cutting on No.2 line extension 0.1km to the east of Fordell Village (NZ38/757839)". However, Milne presents no evidence that the dunesands in the two places are the same unit: the section near Fordell Village does not expose a full section and it is possible that other dunesands are present lower down in the cover bed sequence).

In accord with the dates presented in this work it is suggested that the Mt Curl Tephra should overlie the Brunswick Terrace, and that the dunesand at the Mt Curl type locality is related to the cutting of the Ngarino Terrace. The Mt Curl Tephra has yet to be identified on any marine terrace of the South Taranaki-Wanganui sequence which suggests that it thins rapidly west of its type section so as to make simple visual recognition unlikely. Further work, particularly east of Wanganui is required to properly test the stratigraphic relationship of the Mt Curl Tephra to the South Taranaki Terrace sequence.
# APPENDIX 4

**SURVEYED TERRACE FRONT VALLEYS**

<table>
<thead>
<tr>
<th>Valley Number</th>
<th>Grid Reference of stream exit</th>
<th>Standline</th>
</tr>
</thead>
<tbody>
<tr>
<td>SV-001</td>
<td>N129/976305</td>
<td>Brunswick</td>
</tr>
<tr>
<td>SV-002</td>
<td>N129/976305</td>
<td>Brunswick</td>
</tr>
<tr>
<td>SV-003</td>
<td>N129/981301</td>
<td>Brunswick</td>
</tr>
<tr>
<td>SV-004</td>
<td>N129/920315</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-005</td>
<td>N129/920315</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-006</td>
<td>N129/920315</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-007</td>
<td>N129/920315</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-008</td>
<td>N129/939353</td>
<td>Brunswick</td>
</tr>
<tr>
<td>SV-009</td>
<td>N129/939353</td>
<td>Brunswick</td>
</tr>
<tr>
<td>SV-010</td>
<td>N130/188130</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-011</td>
<td>N130/183131</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-012</td>
<td>N130/185131</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-013</td>
<td>N130/192130</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-014</td>
<td>N130/193130</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-015</td>
<td>N130/234121</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-016</td>
<td>N130/236119</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-017</td>
<td>N130/235120</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-018</td>
<td>N130/237119</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-019</td>
<td>N130/228124</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-020</td>
<td>N130/230123</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-021</td>
<td>N130/229123</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-022</td>
<td>N130/224125</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-023</td>
<td>N130/224125</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-024</td>
<td>N130/232143</td>
<td>Brunswick</td>
</tr>
<tr>
<td>SV-025</td>
<td>N130/195130</td>
<td>Ngarino</td>
</tr>
<tr>
<td>SV-026</td>
<td>N130/182154</td>
<td>Brunswick</td>
</tr>
<tr>
<td>SV-027</td>
<td>N129/012113</td>
<td>Inaha</td>
</tr>
</tbody>
</table>

*NB: Valley long profiles - all figures in metres eg. 25(-6)*

*horizontal distance (vertical fall).*
SV-004

96 (-17)

106 (-16)

39/1

39/2

76 (-6.5)

96 (-4.5)

70 (-3)

88 (-2)

Brunswick Terrace surface

Horizontal & vertical scale
Brunswick Terrace surface
wave cut surface
thalweg

SV-018

101
102
103

55 (-3)
65 (-3)
66 (-3)

28 (-0.9)
27 (-0.9)
16 (-0.9)

metres

Horizontal & vertical scale

50 (-3.5)
37 (-3)

5 (-3.5)

Brunswick Terrace surface

Wave cut surface

Junction

Thalweg

Junction

metres

Horizontal & vertical scale
SV-02: metres Brunswick Terrace surface
wave cut surface
junction
thalweg

Horizontal & vertical scale
SV-023

Brunswick Terrace surface

wave cut surface

thalweg

metres

Horizontal & vertical scale

50

20

10

0


AITKEN, J.F., in press. The occurrence and inferred age of a fifth coastal terrace near Hawera, New Zealand. NZ Soil Bureau Publ.


DAVIS, W.M., 1908. Deflection of rivers by the earth's rotation Science, 27, pp.32-33.


GOW, A.J., 1967. Petrographic studies of iron sands and associated sediments near Hawera, South Taranaki NZ Jnl Geol Geophys, 10, pp.675-695.


GRANGE, L.I., 1927. The geology of the Tongaporutu-Ohura subdivision NZ Geol Surv Bull 31, 63pp.


HADLEY, R.F., 1961. Some effects of microclimate on slope morphology and drainage basin development *US Geol Surv Prof Paper* 424-B.


HIRANO, M., 1975. Simulation of developmental process of interfluvial slopes with reference to graded form *J Geol*, 83, pp.113-123.


MOORE, P.R., 1975. Measuring flow of small streams : use of a portable weir Tane, 21, pp.147-152.


PELTIER, L.C., 1950. The geomorphic cycle in periglacial regions as it is related to climatic geomorphology Ann Assoc Amer Geogr, 40, pp.214-236.


ROBERTSON, N.G., 1963. The frequency of high intensity rainfall in New Zealand NZ Met S Misc Publ 118.


SEWARD, D., 1979. Comparison of zircon and glass fission-track ages from tephra horizons Geology, 7, pp.479-482.


SHULITS, S., 1941. Rational equation of River bed profile Am Geophys Union Trans, 36, pp.655-663.


TE PUNGA, M.T., 1952a. The geology of Rangitikei Valley Memoir NZ Geol Survey 8, 44pp.

TE PUNGA, M.T., 1957b. A conformable sequence of rocks of the Wanganui and the Hawera Series NZ J Sci & Technol, B38, pp.328-341.


ZINGG, A.W., 1940. Degree and length of land slope as it affects soil loss in runoff, Agric Eng, 21, pp.59-64.