THE UNIVERSITY OF WAIKATO
Te Whare Wānanga o Waikato
Research Commons

http://researchcommons.waikato.ac.nz/

Research Commons at the University of Waikato

Copyright Statement:

The digital copy of this thesis is protected by the Copyright Act 1994 (New Zealand).

The thesis may be consulted by you, provided you comply with the provisions of the Act and the following conditions of use:

- Any use you make of these documents or images must be for research or private study purposes only, and you may not make them available to any other person.
- Authors control the copyright of their thesis. You will recognise the author's right to be identified as the author of the thesis, and due acknowledgement will be made to the author where appropriate.
- You will obtain the author's permission before publishing any material from the thesis.

Characteristics and evolution of a dynamic prograding continental margin: the late Neogene Giant Foresets Formation, northern Taranaki Basin, New Zealand.

A thesis submitted in fulfilment
of the requirements for the Degree of

Doctor of Philosophy in Earth Sciences
at the
University of Waikato

by

ROCHELLE JANICE HANSEN

2003

The University of Waikato
Te Whare Wananga o Waikato

The University of Waikato
Hamilton, New Zealand

Abstract

The Pliocene-Pleistocene Giant Foresets Formation represents the progradation and development of the modern continental margin in northern Taranaki Basin. The dynamic evolution of this formation is fundamentally linked to a Pliocene-Pleistocene increase in sediment flux sourced from South Island, to extensional graben formation in northern Taranaki Basin, to flexural subsidence in Wanganui Basin, and to uplift and erosion in King Country Basin. The development of the Giant Foresets Formation is summarised in this thesis by a series of paleogeographic maps based on the interpretation of wireline data from hydrocarbon exploration holes, industry-acquired seismic reflection data, and the paleoenvironmental interpretation of foraminiferal data derived from well cuttings samples.

Outbuilding of the modern continental shelf-slope margin began during the Early Miocene as the regressive phase of a 1st-order megacycle. Four 2nd-order cycles superimposed on this megacycle have been identified in Taranaki, Wanganui and King Country Basins. The evolution of the Giant Foresets Formation is closely linked to the youngest two megacycles: the Mid Miocene-earliest Pliocene Whangamomona Group/megacycle, and the mid Pliocene-Pleistocene Rangitikei Supergroup/megacycle. The Whangamomona Group includes the Otunui, Mount Messenger, Urenui and Kiore Formations. The continental margin associated with the Whangamomona Group had only limited extent in southern and eastern parts of Taranaki Basin. The Rangitikei Supergroup in Wanganui Basin includes the Tangahoe Mudstone and younger Plio-Pleistocene sediments. The Giant Foresets Formation represents the Rangitikei Supergroup in northern Taranaki Basin.

Revision of the biostratigraphy of four well sections, and integration with biostratigraphic data from other well sections, has highlighted the development in northern Taranaki Basin of a region-wide unconformity between Miocene (Late Tongaporutuan to Late Kapitean) and Pliocene (Early to mid Opoitian) strata, prior to deposition of the Giant Foresets Formation. This is expressed as a paraconformity across much of the northern part of the Western Stable Platform (Ariki Formation), thinner condensed intervals in the Northern Graben, and an erosional unconformity across the Turi Fault Zone. The Ariki Formation is related to at least three important events: (i) the limited extent of progradation of the Whangamomona Group foreset front; (ii) initiation of extension of the Northern Graben; and (iii), the timing of the Tangahoe pulldown in Wanganui Basin. Palinspastic restoration of a complete seismic reflection profile, and backstripping and decompaction of eleven well sections, suggests that paleobathymetric relief may have been locally important in extending the duration of terrigenous sediment starvation.

Six wireline facies have been identified in the Giant Foresets Formation and underlying units. These are: (i) hemipelagic facies; (ii) basin floor fan facies; (iii) slope fan facies; (iv) shelf facies; (v) slumped facies; and (vi) condensed facies. A number of subfacies have also been identified. Construction of a fence diagram using the facies motifs identified in this study has highlighted the distribution of volcaniclastic versus siliciclastic and hemipelagic sedimentation patterns in both time and space in northern Taranaki Basin. The Miocene was dominated by hemipelagic facies interfingering with coarser lithologies sourced predominantly from reworked volcaniclastic material, which travelled some distance into the northwestern region of northern Taranaki Basin. The Plio-Pleistocene Giant Foresets Formation is characterised by a hemipelagic motif, reflecting the dominantly fine-grained nature of the unit, although the formation displays an overall coarsening-upward trend, reflecting increasing proximity to the advancing foreset front and the development of shelf deposits. Localised Pliocene basin floor fans in the Northern Graben (Mangaa Formation) emphasize the role of the graben as a sink for sedimentation during the Opoitian and Waipipian.

Identification and mapping of more than 70 seismic units within the Giant Foresets Formation in northern Taranaki Basin has allowed the post-Miocene sediment distribution patterns to be characterised. Isopach maps and structure contour maps demonstrate the translation of depocentres through time, and illustrate how the sediment distribution pattern associated with the advancing continental margin gradually changed from one of focused sedimentation in the south, and along the axis of the Northern Graben (Mangaa Formation), to a more linear front. Numerous channels and channel systems have been mapped in the Mangapanian to Nukumaruan shelf succession, often preferentially oriented along the axis of the Northern Graben.

The muddy texture of the Giant Foresets Formation is also shown by the seismic characteristics of the formation, with architectural elements and depositional styles reflecting mud- and mixed mud/sand-rich depositional systems. Basin floor (bottomset) facies are characterised by low- relief mounds and/or gull-wing geometries of channel-levee systems, or sub-parallel reflectors, while slope (foreset) facies are dominated by chaotic reflection configurations consistent with slumped deposits or numerous stacked small-scale channel-levee systems on the middle and lower slope.

Much of the shallowing of water depth in northern Taranaki basin occurred during deposition of bottomset facies. Depositional environments changed from mid/lower bathyal during the Late Pliocene, to upper bathyal during the Waipipian and Mangapanian. However, the most rapid aggradation and progradation of the continental margin occurred during the Mangapanian to Late Nukumaruan, with progradational rates as high as 35 km/m.y. recorded. While the coarsest sediment textures are associated with shelf (topset) facies, and the youngest (degradational) foreset units, examination of benthic foraminiferal samples indicate that there was effective transport from the shelf to the base of slope throughout the depositional history of the Giant Foresets Formation.

A dramatic (uphole) decrease in planktic foraminiferal percentage between the Ariki/ Manganui Formations and the Giant Foresets Formation, with a corresponding change in watermass conditions but no dramatic change in water depth, suggests that the progradation of the continental margin during the latest Miocene and Early Pliocene acted to deflect coastal upwelling cells away from northern Taranaki Basin, decreasing available surface water nutrients, and thus reducing the abundance of planktic foraminifera during this time.

Cyclicity within the Giant Foresets Formation is evident on seismic reflection profiles, in wireline motifs, and from paleontological evidence. Sequences may be defined by a particular wireline motif, or set of motifs, the use of seismic reflection configurations to identify components of depositional environments interpreted to represent regressive (RST) and lowstand (LST) sea level conditions, and through the identification of mixed deep water and shallow water benthic foraminiferal assemblages. Basinward of the shelf margin, the highstand systems tract (HST) is often only represented by a bright, high amplitude and laterally continuous reflector that rolls over into shelf deposits. Transgressive (TST) and HST components of a relative sea level cycle cannot be separated from each other on seismic reflection profiles of the shelf deposits of the Giant Foresets Formation. Sequence boundaries often correspond to a bold seismic reflector horizon. A number of discrete high order sequences have been identified, but many of these cannot be consistently correlated between well sections. However, three levels of cyclicity can be established: 2nd-order cycles related to progradation of the Whangamomona Group and Rangitikei Supergroup, 4th-order (400 ka) cycles, and ?5th- order (100 ka) cycles (possibly related to glacio-eustatic controls). Sequences are highly asymmetrical, with most of the deposition occurring during lowering and low sea level conditions. Sediment flux is the predominant factor controlling the seismic character of well sections. However, three levels of cyclicity can be established: 2^{nd}-order cycles related to progradation of the Whangamomona Group and Rangitikei Supergroup, 4^{th}-order (400 ka) cycles, and ?5^{th}- order (100 ka) cycles (possibly related to glacio-eustatic controls). Sequences are highly asymmetrical, with most of the deposition occurring during lowering and low sea level conditions. Sediment flux is the predominant factor controlling the seismic character of the Late Opoitian to Mangapanian parts of the Giant Foresets Formation, while changes in relative sea level during the Nukumaruan to Castlecliffian increasingly influenced stratal architecture. The shoreline is not considered to have moved basinward of the shelf break during any of the sea level cycles represented in the Giant Foresets Formation, except possibly for a few cycles during the Late Nukumaruan and Castlecliffian.

Acknowledgements

It is the nature of a project such as this that it will only evolve and be accomplished with the help of a large and varied group of people. Thus there are a number of people who can also take credit for this work finally being completed.

The biggest thank you must go to Professor Peter Kamp, who first suggested that I do a PhD at a time when I really needed a new direction and focus, and who has supported this research and offered unwavering encouragement and advice, and who has always believed in my academic abilities. I would also like to acknowledge the scholarship and research funding provided through the Energy Resources research programme funded by FRST and led by Peter Kamp. I also give grateful thanks to Professor Cam Nelson for helping to give me a background in sedimentary geology, for discussion about my work, and for editing some of my chapters.

A special thanks goes to (Dr.) Penny Cooke, for the huge effort she put in during the final weeks in helping me to get this project finished, the friendship offered over the last five years, and the emotional encouragement when things were looking grim. To the numerous other PhD students who kept me going, including Kyle, Adam, Arne, Steve (Dr.), Avon (Dr.), and Vincent (Dr.) who offered humour, support, technical advice, and companionship on the soccer field (pity we never won), I offer you all a huge thank you.

I am extremely grateful and appreciative of the help provided by Martin Crundwell (IGNS), who gave me assistance with the identification of foraminiferal samples. Also to Dr. Bruce Hayward (Private Science Provider) who helped edit my paleoenvironmental chapter, and offered excellent suggestions and advice, and Dr. Peter King and George Scott (GNS) for willingly answering my numerous questions about Taranaki Basin and Neogene foraminifera. To the team at Geosphere Exploration Ltd., particularly Dr. Glen Thrasher, Dr. Mac Beggs, and Dr. Beate Leitner, I offer my thanks for the provision of office space, for tutoring me in the fundamentals of seismic interpretation, and encouragement with my work, and for access to some of your data.

BSc Tech students Daniel Butler and (especially) Candice Bardsley will always gratefully be remembered for the huge effort they put in over their summer 'breaks' to process data and complete much of the time-consuming and repetitive work that needed to be done in order to finish this thesis. A general thanks is also extended to the Department of Earth Sciences technical staff and other departmental staff who over the years have offered help, answered questions, and provided technical support when needed, and to Daryl Gillgren (formerly of the Geography Department, University of Waikato) who wrote the programmes needed for conversion of paper seismic traces to digital format.

Finally, to my much loved family, especially my mum and dad, my numerous very good friends (you know who you are), my long suffering husband John, and our unknown factor - thank God, its finally finished. The beast is dead!

Table of Contents

Chapter 1: Introduction

Chapter 1: Introduction

1.1 Overview

Being New Zealand's only commercially proven hydrocarbon sedimentary basin, Taranaki Basin (Fig. 1.1) has been the focus of numerous petroleum-oriented investigations since the early 1900s, when a small sub-commercial oil field was discovered near New Plymouth (Bennett et al., 1992). Many wells have since been drilled, predominantly in the southern offshore region and onshore Taranaki Peninsula. Most wells have targeted Paleogene sandstone and limestone units, although more recent wells have discovered hydrocarbon accumulations in Miocene to earliest Pliocene sandstone of the Moki, Mount Messenger and Matemateaonga Formations (refer to Chapter 2 for lithostratigraphy). Noticeably fewer wells have been drilled in the northern offshore region of Taranaki Basin. In this region, other than a few noteworthy shows in a handful of wells, sub-commercial accumulations have only been discovered in volcaniclastic sediments associated with Miocene volcanic complexes (ARCO Petroleum NZ Inc., 1988). This latter find initiated a surge of exploration drilling during the late 1980s and early 1990s, including the acquisition of seismic reflection data, and through the Cretaceous-Cenozoic Project (King and Thrasher, 1996), the biostratigraphic re-evaluation of a number of earlier well sections.

Although exploration results indicate that a hydrocarbon system has been active in northern Taranaki Basin, all wells drilled to date have subsequently been plugged and abandoned, most as dry holes. Because of this, interest in the northern region has waned significantly, and as a consequence much less is known about its evolution and depositional history than southern and onshore regions. While exceptional outcrops of Miocene to Recent sediments occur on the well-exposed northern Taranaki coastline, affording excellent opportunities for detailed investigation, the same obviously cannot be said for the offshore northern region. Pliocene and Pleistocene sediments were never a target during the drilling process in this region, and thus little attention was paid to the sedimentary characteristics and information that could be imparted from geophysical and paleoenvironmental data. However, one particular characteristic of the Plio-Pleistocene to Recent sedimentary column *has* aroused interest for several decades – the spectacular stacked succession of bold, off-lapping, sinusoidal clinoforms that underlie the shelf and prograde toward the modern continental slope, as observed on seismic reflection profiles (see Chapter 2). This seismic characteristic has given rise to the name by which this sequence of sediments is known: the Giant Foresets Formation.

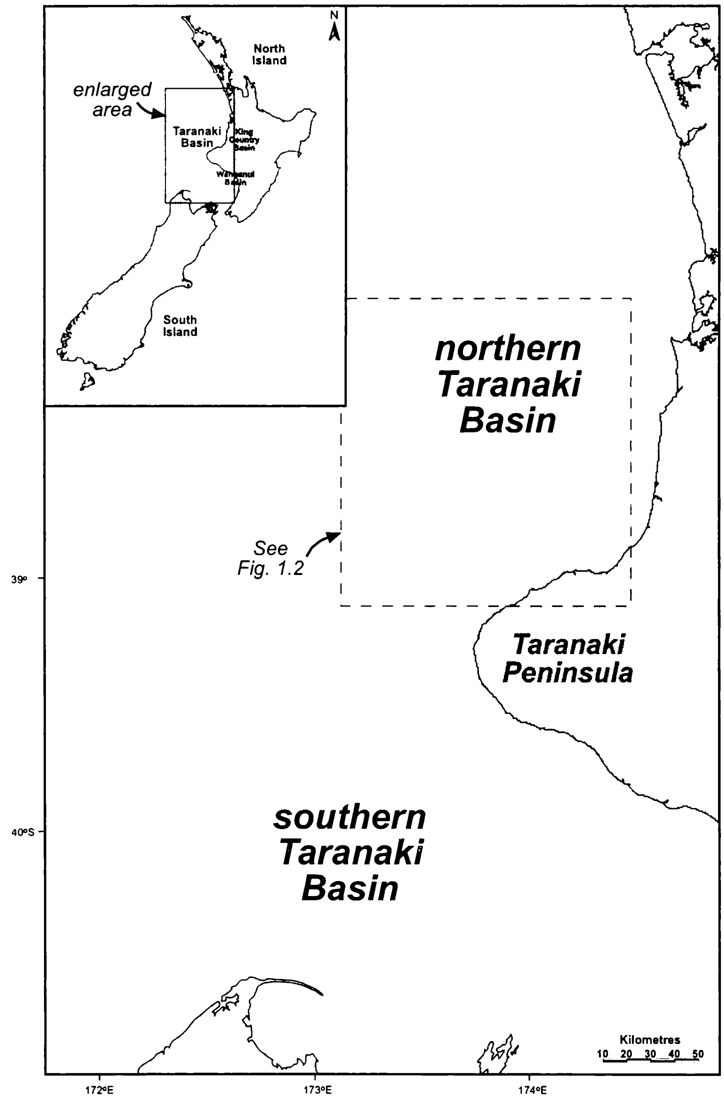


Fig. 1.1: Location of Taranaki Basin in New Zealand, showing broad northern and southern Taranaki Basin subdivisions. Position of Wanganui Basin and King Country Basin also shown in inset.

The Giant Foresets Formation was first used in a stratigraphic context by Shell BP Todd (1977), to describe a 2000 m-thick, seismically distinctive fine-grained sequence encountered in Tane-1 (Fig. 1.2). The characteristic clinoform seismic character of the Giant Foresets Formation records the westward and northward growth of the continental shelf, a process which began in the early Miocene with the initiation of uplift and erosion of basement following the development of the Australia-Pacific plate boundary (Alpine Fault-Hikurangi Subduction Zone) as a tectonic feature through New Zealand. It is an extremely thick deposit, in places reaching more than 2 km, despite the fact that it encompasses no more than 5 m.y. It is thickest in the northern part of Taranaki Basin, particularly over the Western Stable Platform and within the Northern Graben (Chapter 2), and embraces some of the highest sedimentation rates in northern Taranaki Basin (up to 1.0 m/ka; Beggs, 1990). Laterally equivalent deposits include Plio-Pleistocene basin floor to shelf sediments (Rangitikei Supergroup) studied in detail in the Wanganui Basin (e.g., Naish and Kamp, 1994; McIntyre, 2001; Kamp et al., 2002; Vonk et al., 2002).

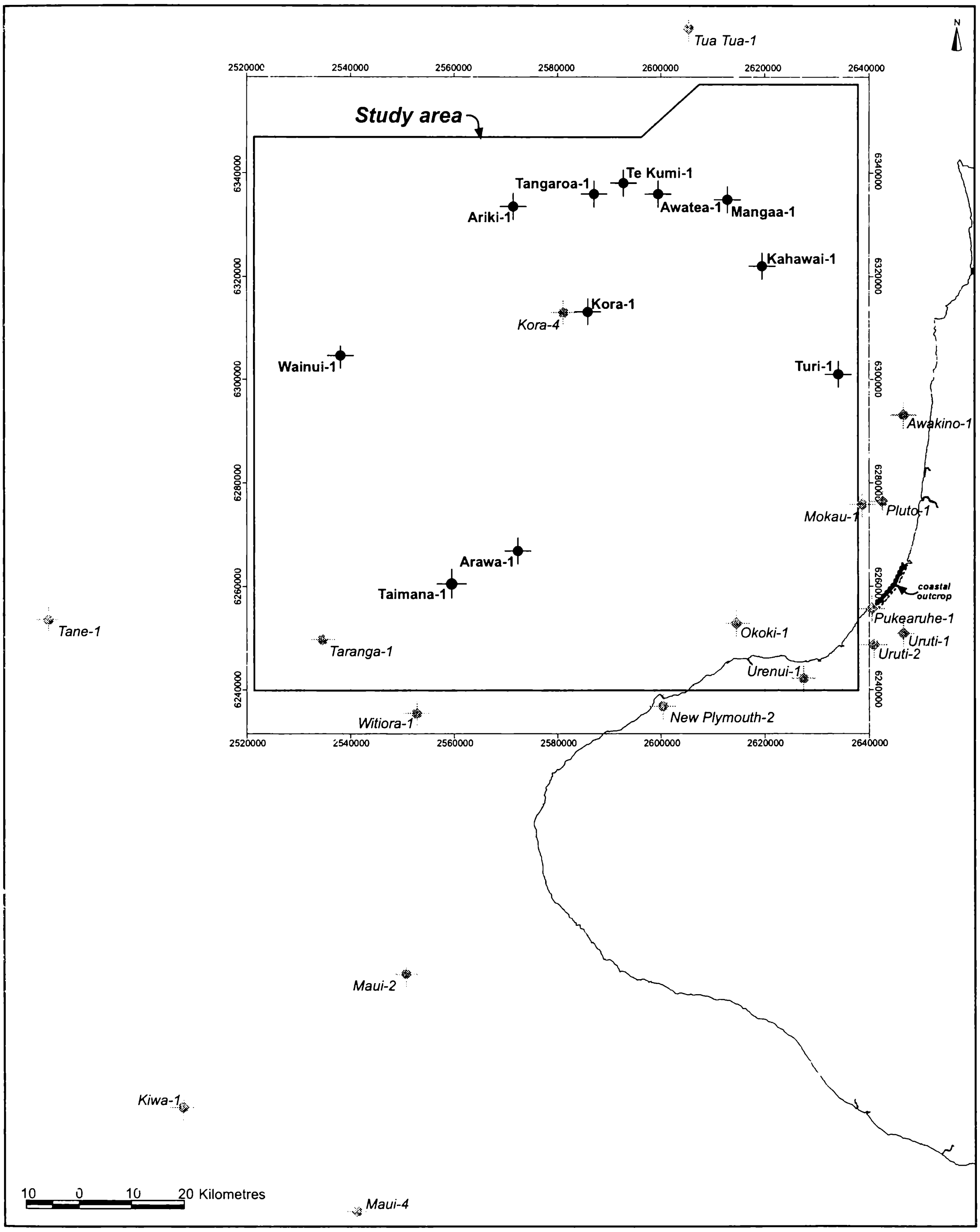


Fig. 1.2: Location map illustrating the approximate extent of the area covered by this study. Primary wells utilised in this study are shown as black well markers and bold names. Other wells of interest are indicated by lighter shade well markers and italicised well names. Small shaded box indicates the coastline from which outcrop analogues were obtained. New Zealand map grid projection for the study area is also displayed.

Somewhat surprisingly, given its prominence in the stratigraphic record, no previous studies have investigated in detail the depositional evolution of this formation. Previously, Beggs (1990) looked at the gross seismic characteristics of the Giant Foresets Formation, and recognised four distinct seismic components. Soenander (1991, 1992) mapped four major depositional units (sequences) within the Giant Foresets Formation, each separated by a prominent reflector and characterised by specific internal seismic reflector characteristics. That study highlighted the progradational direction of the foresets and concluded that clinoforms prograded in the form of migrating fan lobes. Other studies have used foreset clinoforms to calibrate the paleobathymetric position of benthic foraminifera by extrapolating present day water depths to the modern day shelf break (e.g., Hayward, 1990a; Crundwell et al., 1994). Ogilvie (1993) discussed the influence of eustasy, subsidence and sedimentation on the distribution of sediments within the Giant Foresets Formation. In the last few years some studies have considered the impact that this formation has had upon the thermal regime in Taranaki Basin, the maturation of underlying units, and hydrocarbon migration (e.g., McAlpine, 2000; Stagpoole, 2000).

This PhD study builds upon previous investigations by integrating several different techniques, including biostratigraphic dating and paleoenvironmental analysis from planktic and benthic foraminiferal assemblages, analysis and interpretation of geophysical (wireline) logs and seismic reflection profiles, the development of outcrop analogues from the coastal north Taranaki section, and decompaction and backstripping for palinspastic reconstruction. It is intended that this thesis provide a framework upon which even more detailed investigations can build in the future.

The primary study area covers the greater part of the offshore northern Taranaki Basin (Fig. 1.2), encompassing an area of approximately 12,000 km^2. Eleven primary wells, identified in Fig. 1.2, are used to provide biostratigraphic and lithostratigraphic control within the study area, although information from other wells is used where appropriate. Correlation between wells is achieved primarily through the use of seismic reflection mapping.

1.2 Objectives of study

The primary objective of this study is to understand and document the Plio-Pleistocene evolution of the Giant Foresets Formation in northern Taranaki Basin. With this is mind, and with the resources available to this study, several secondary objectives have been identified:

1. To review the chronology and paleoenvironmental characteristics of the Giant Foresets Formation by reassessing four key wells, and where possible to apply this information to other wells within the study area for which there are published records.
2. To document the seismic and wireline characteristics of the Giant Foresets Formation and to investigate the link with depositional paleoenvironments and lithological characteristics.
3. To document the depositional trends associated with progradation of the continental margin by interpreting and mapping the distribution of discrete seismic packages.
4. To understand the cyclic nature of the formation, the degree (order) of cyclicity that can be recognised, and the wireline, seismic and biostratigraphical characteristics of cycles.
5. To document the impact that major structural features (faults, volcanic centres) have had on depositional patterns across the study area.
6. To complete backstripping and decompaction of an entire seismic reflection profile for palinspastic reconstruction to understand more fully the evolution of the basin.
7. To review the literature regarding the hydrocarbon prospectivity of the northern Taranaki Basin and to assess the role that the Giant Foresets Formation has played on the petroleum system.
8. To illustrate the Plio-Pleistocene evolution of northern Taranaki Basin as a series of paleogeographic maps.

1.3 Thesis outline

A systematic approach to each of the major topics (chronology, wireline and seismic characteristics, paleoenvironmental analysis, and palinspastic reconstruction) has been undertaken in this thesis. Each chapter introduces and expands upon a new stratigraphic tool, although elements from previous or subsequent chapters are included to emphasise ideas or concepts and to provide an integrated account. While the concept of cyclicity may warrant individual chapter status, it is instead treated slightly differently, being developed through Chapters 4 and 5, before being discussed more fully in Chapter 6. The following outlines the contents of each chapter.

Volume 1

Chapter 1 - Introduction

This chapter provides an overview of the study, a brief introduction to the study area, and discussion of the objectives of the study, as well as presentation of the structure of the thesis.

Chapter 2 - Structural evolution and lithostratigraphy of Taranaki Basin, with emphasis on offshore northern Taranaki Basin

This chapter presents a literature review of the structural evolution and lithostratigraphy of Taranaki Basin, with emphasis on post-Miocene strata in the northern region. A discussion of the stratigraphic links between Wanganui and Taranaki Basins are also given.

Chapter 3 - Biostratigraphic review of northern Taranaki Basin

This chapter reassesses the biostratigraphy of four key wells within the study area, revising or constraining the stratigraphic position of New Zealand Stage boundaries. The implications are considered for the biostratigraphic dating of the Giant Foresets Formation intersected in other well sections. This chapter also discusses the nature and chronostratigraphic significance of the Ariki Formation and its importance as a prelude to accumulation of the Giant Foresets Formation.

Chapter 4 - Geophysical wireline log characterisation of late Neogene sediments, offshore northern Taranaki

Wireline logging interpretative methods are applied to the latest Miocene-Pleistocene succession intercepted in wells drilled within the study area. In conjunction with well cuttings and core descriptions from well completion reports, the subsurface characteristics of various late Neogene formations, particularly the Giant Foresets Formation, are described and compared to onshore (outcrop) analogues. This has allowed the identification and development of discrete wireline facies, each of which document a specific mode of deposition and depositional environment. The relationship between Taranaki Basin and Wanganui Basin is briefly examined, and this has assisted the recognition of cyclical packages based on wireline signatures within the Giant Foresets Formation.

Chapter 5 - Seismic geometry and sequence architecture of the Giant Foresets Formation, with emphasis on the Northern Graben region.

A total of 32 seismic lines have been interpreted and more than 70 discrete seismic packages identified and mapped over a wide area. This chapter outlines the methodology behind the interpretation of these lines, and the results obtained. Structure contour and isopach maps allow the extent of shelf progradation to be mapped through time, the changing direction of progradation to be observed, and depositional loci to be identified. The influence of major structural features and tectonic events on sediment distribution is also discussed. Seismic facies have been identified and described, and these may later help to create much more detailed time slices in future projects. The cyclothemic nature of the Giant Foresets Formation is discussed in this chapter, and the concept of cyclicity is further developed.

Chapter 6 - Paleoenvironmental interpretation of four wells in northern Taranaki Basin: implications for paleogeographic reconstructions

This chapter investigates the paleoenvironmental signatures obtained from the detailed re-evaluation of four key well sections in the study area. Paleobathymetric curves and water mass conditions are derived from trends in benthic and planktic foraminifera, while foraminiferal abundance peaks and various statistical analyses of benthic species are interpreted to distinguish other environmental conditions. Cyclicity within the Giant Foresets Formation is developed to its fullest extent with the integration of wireline and seismic data. This information is incorporated into a model to better understand the cyclothemic nature of the formation as imaged on seismic reflection profiles

Chapter 7 - Dynamic evolution of a prograding continental margin

Backstripping and decompaction of individual wells, and palinspastic restoration of a complete seismic reflection profile, coupled with paleogeographic maps, are utilised in this chapter to summarise the geological evolution of the Giant Foresets Formation. A summation of the petroleum systems and hydrocarbon prospectivity of northern Taranaki Basin is also given, with a brief discourse on the role that the Giant Foresets Formation has played in maturation and migration of hydrocarbons.

Chapter 8 - Summary and conclusions

This chapter provides a brief summary of the main concepts and information generated and presented in successive chapters of the thesis.

<u>Volume 2</u>

Appendices and Enclosures.

Chapter 2: Review of the structural evolution and lithostratigraphy of Taranaki Basin, with emphasis on offshore northern Taranaki Basin

Chapter 2: Review of the structural evolution and lithostratigraphy of Taranaki Basin, with emphasis on offshore northern Taranaki Basin

2.1 Introduction

Hydrocarbons were first identified as a significant resource in New Zealand in 1959, with the discovery of gas-condensate at Kapuni, on the Taranaki Peninsula, and in 1969 with the offshore discovery of the Maui gas-condensate field. Over 70% of New Zealand's hydrocarbon production comes from these two fields, and as New Zealand's only economic hydrocarbon-producing sedimentary basin, Taranaki Basin has been the focus of extensive research and exploration during the latter half of the 20th Century. Much literature has been published in an effort to understand the structural evolution, stratigraphy, lithology, and petroleum potential of the greater Taranaki Basin, culminating in the monograph (King and Thrasher, 1996) published by the Institute of Geological and Nuclear Sciences Ltd. While this monograph presents an excellent synthesis of Taranaki Basin's geology and petroleum systems, there is scope for much more detailed work, such as that attempted in this PhD study.

This chapter presents a broad literature review of the Cretaceous-Cenozoic evolutionary history and stratigraphy of Taranaki Basin, focusing in particular on the Late Neogene to Recent, and with emphasis on northern (offshore) Taranaki Basin. For a more comprehensive lithostratigraphic review, the reader is directed towards King and Thrasher (1996), and King (2000a,b).

2.2 Tectonic setting and subsurface architecture

Taranaki Basin presently lies in a back-arc position approximately 400-500 km behind the Hikurangi Trough, which is the structural trench (Bennett et al., 1992; Stern and Davey, 1990; Fig. 2.1). The basin is located along the western side of the North Island, and is mainly offshore. Taranaki Basin morphology is bounded by several major structural and tectonic features (Fig. 2.2a,b). To the east, the north-south trending Taranaki Fault has been an important structural control on the development of the basin, particularly through the Early Miocene. North of the Taranaki Peninsula, the Taranaki Fault occurs 6-10 km offshore, running sub-parallel to the modern shoreline before passing under the peninsula (Fig. 2.2a). The Taranaki Fault bounds part of the Southern Inversion Zone, and marks the western margin of the Mesozoic Murihiku basement terrane exposed onshore as the Kawhia Syncline. The Patea-Tongaporutu High, delineated via gravity anomalies in the subsurface (e.g., Mills, 1989), lies to the east of the fault. Together these basement structural highs delineate the eastern margin of Taranaki Basin and separate it from the more easterly-located Wanganui and King Country Basins (Mills, 1989; Thrasher, 1990; Bennett et al., 1992; King et al., 1993; King and Thrasher, 1996; Kamp et al., 2002). To the west, Taranaki Basin steps up on to the Challenger Plateau (Thrasher, 1990). Both northern and southern limits are not well defined, due to lack of seismic and well control, although the northern limit is arbitrarily cut off at 38°S, and the southern extent includes the northwestern South Island (approx. latitude 41°30'S; King and Thrasher, 1996). Middle-to-late Miocene andesitic volcanic edifices (now completely buried) are aligned on a north-northeast-south-southwest trend (King and Thrasher, 1996), sub-parallel to the modern northern Taranaki coastline.

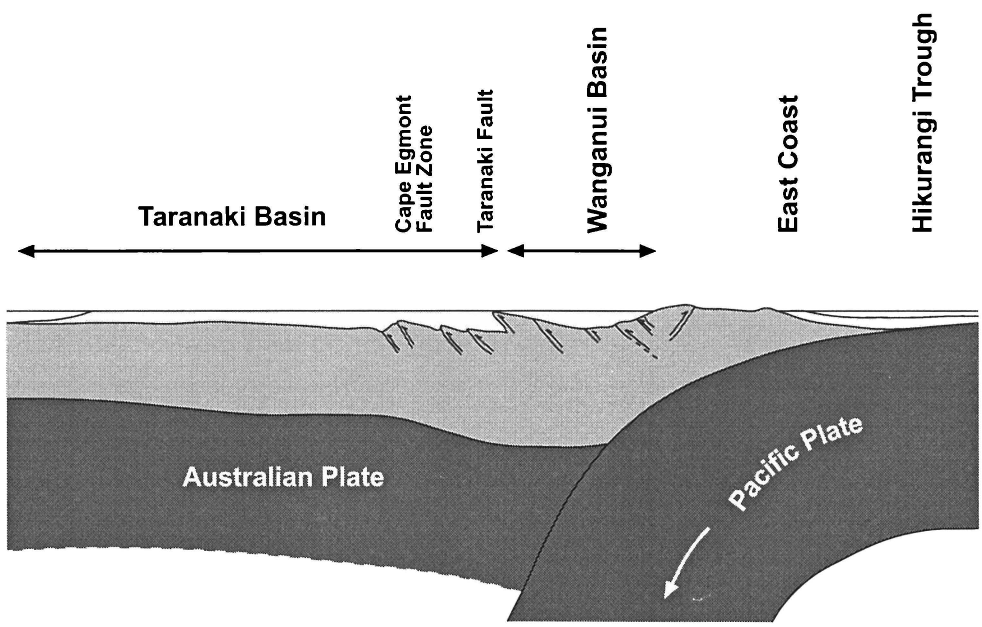


Fig. 2.1: Schematic NW-SE cross section across the present day plate boundary zone, illustrating the back-arc position of Taranaki Basin. Adapted from Holt and Stern, 1994.

Pilaar and Wakefield (1978) identified two structural units within the basin: a stable platform to the west (Western Platform) and a graben complex to the east (Taranaki Graben complex). Knox (1982) identified similar, though not identical, units, subsequently labelled Western Unit, Southern Unit and Northern Unit. King and Thrasher (1992) subdivided the basin into 5 main structural provinces. These comprised the Western Platform, North Taranaki Graben, South Taranaki Graben, Tarata Thrust Zone, and Southern Inversion Zone. Subsequent name changes involved the North Taranaki Graben changing to Northern Graben, while the South Taranaki

Graben was renamed the Central Graben, and the Western Platform became the Western Stable Platform. Collectively, all provinces other than the Western Stable Platform fall in the structural region termed the Eastern Mobile Belt (King and Thrasher, 1996). The Eastern Mobile Belt has a composite architecture and evolution, and has been variably uplifted and eroded (Armstrong et al., 1998), while the Western Stable Platform has remained relatively quiescent since Cretaceous times (King and Thrasher, 1996). Of particular interest to this study is the Northern Graben, which is a northwest-southeast trending graben that formed during the latest Miocene and Pliocene. It is bound to the east by the Turi Fault Zone, and to the west by the 30-km wide Cape Egmont Fault Zone. In both fault zones, major faults are still considered to be active. In southern Taranaki Basin, sediment offset on the Cape Egmont Fault Zone is observed to continue through to the present day, displacing the modern seabed up to 5 m in places (Nodder, 1993).

2.3 Polyphase evolution of Taranaki Basin

2.3.1 Broad Cenozoic evolution

Taranaki Basin evolved during the late Cretaceous – Cenozoic through five distinct phases related to changing tectonic regimes. These include rift transform, passive margin, wrench transform, back-arc foreland basin, and outer convergent margin phases (King, 1990). Other tectonic elements include platform subsidence related to incipient subduction, and arc volcanism (King and Thrasher, 1996). Each of these phases, illustrated in Fig. 2.3, has contributed to the sedimentary fill and stratigraphic architecture of the basin.

Taranaki Basin evolved from a failed oblique rift in the late Cretaceous into a relatively stable early Cenozoic passive margin (Thrasher, 1990; Bennett et al., 1992; King and Thrasher, 1992). At about 40 Ma, development of the Australia-Pacific plate boundary system through New Zealand started (King, 2000b). Regional subsidence continued throughout the Eocene and into the Oligocene, but this accelerated along the eastern margin of the basin during the Late Oligocene, attributed to platform subsidence due to far field subduction affects (Holt and Stern, 1994). Concomitant uplift and tilting in the vicinity of the Herangi High, commencing in the Late Whaingaroan and culminating in the Late Waitakian (Late Oligocene) has been attributed to initiation of basement overthrusting from the east along the Taranaki Fault (Nelson et al., 1994), in response to the beginning of a new compressive tectonic regime (King, 1990). This resulted in the formation of a foredeep in eastern Taranaki Basin (Knox, 1982; King, 1990; King and Thrasher, 1992; King and Thrasher, 1996), which accumulated mixed carbonate-siliciclastic sediments.

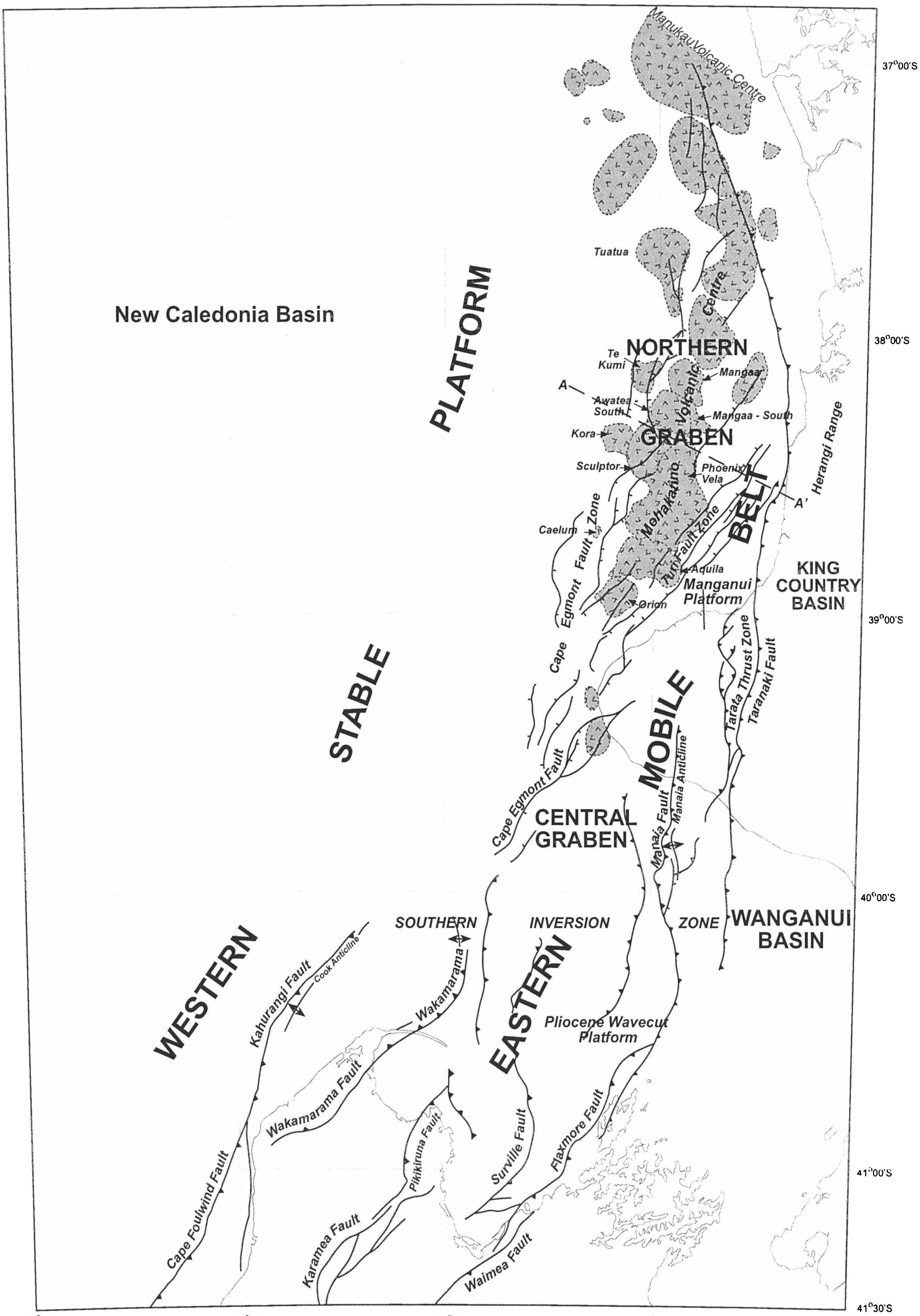


Fig. 2.2a: Structural domains and principal tectonic features, Taranaki Basin. See Fig. 2.2b for northern Taranaki Basin cross section (transect A-A'). Major volcanic centres in the Northern Graben are named on the map. Figure after King et al. (1993), King and Thrasher (1996) and Thrasher et al. (2002). King Country Basin after Kamp et al. (2002).

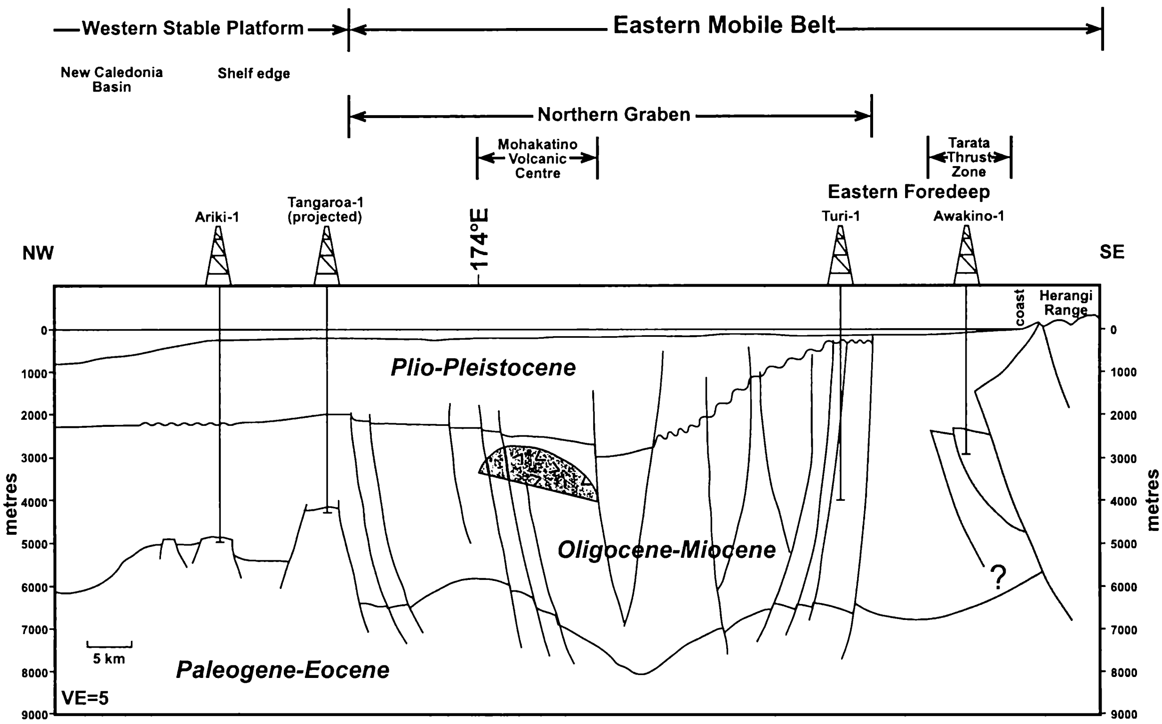


Fig. 2.2b: Cross section across northern Taranaki Basin, illustrating the extent of the main structural and tectonic zones in relation to a structural cross section (Fig. 2.2a) (after King et al., 1993).

The Late Oligocene compressive tectonic regime corresponds to a time of proto-Australia-Pacific plate boundary development through the New Zealand microcontinent (King, 1990; Stern and Davey, 1990; Nelson et al., 1994). Continued southward development and rotation of the Hikurangi subduction zone through the Early Miocene (Otaian) and propagation of dextral transform faulting along the Alpine Fault heralded the onset of convergent tectonics, uplift, expansion of land areas and increased terrigenous sedimentation (King, 2000a). Compression along the eastern margin of Taranaki Basin continued until around 10-12 Ma, with up to 6 km of vertical displacement estimated on the Taranaki Fault (King and Thrasher, 1996). During this time, changes in the plate boundary configuration shifted the focus of compression from the eastern margin southward to the Southern Inversion Zone (King and Thrasher, 1992; 1996), while to the north the Northern Graben formed through extension.

Submarine volcanism, a manifestation of subduction of the oceanic Pacific Plate beneath the Australian plate (King, 1992), occurred during the Middle to Late Miocene. Volcanism appears to have migrated in a southwards direction with time, ultimately leading to the Quaternary volcanics on Taranaki Peninsula (King, 1990). Extension during the Pliocene led to formation of the Turi Fault Zone. This was associated with rifting in the Northern Graben, which was later filled with sediment prograding from the south.

2.3.2 Neogene evolution of northern Taranaki Basin

This section reviews specific Neogene events or features that have contributed to the geometry and architecture of the Miocene to Recent sedimentary basin fill in the northern part of Taranaki Basin. The Neogene evolution is summrised in Fig. 2.3 (from King, 2000a).

2.3.2(i) Miocene (c. 24 Ma – 5.2 Ma)

The early Miocene was characterised by deformation along Taranaki Basin's eastern margin, when basement was thrust westwards across the Taranaki Fault, and foredeep subsidence to the west of the fault (Bennett et al., 1992; King and Thrasher, 1996). On seismic reflection profiles, a basement wedge is imaged overthrusting a Late Oligocene horizon (Tikorangi Formation), and this is onlapped by Early Miocene (c. 22-16 Ma) deposits and buried by younger bathyal sediments (Fig. 2.4). Water depths in the northern part of the basin were possibly greater at this time than at any other time, and obtained a maximum of 1300-1500 m (lower bathyal) at Tangaroa-1 during the Early to Late Miocene (Shell BP and Todd Oil Services Ltd., 1981). It was not until the late-Early Miocene, long the basin's southern and southeastern margins, that the regressional phase started, and sediment supply rates began to exceed subsidence rates. This marked the beginning of the aggradation of the basin floor, and northward progradation of a continental margin wedge.

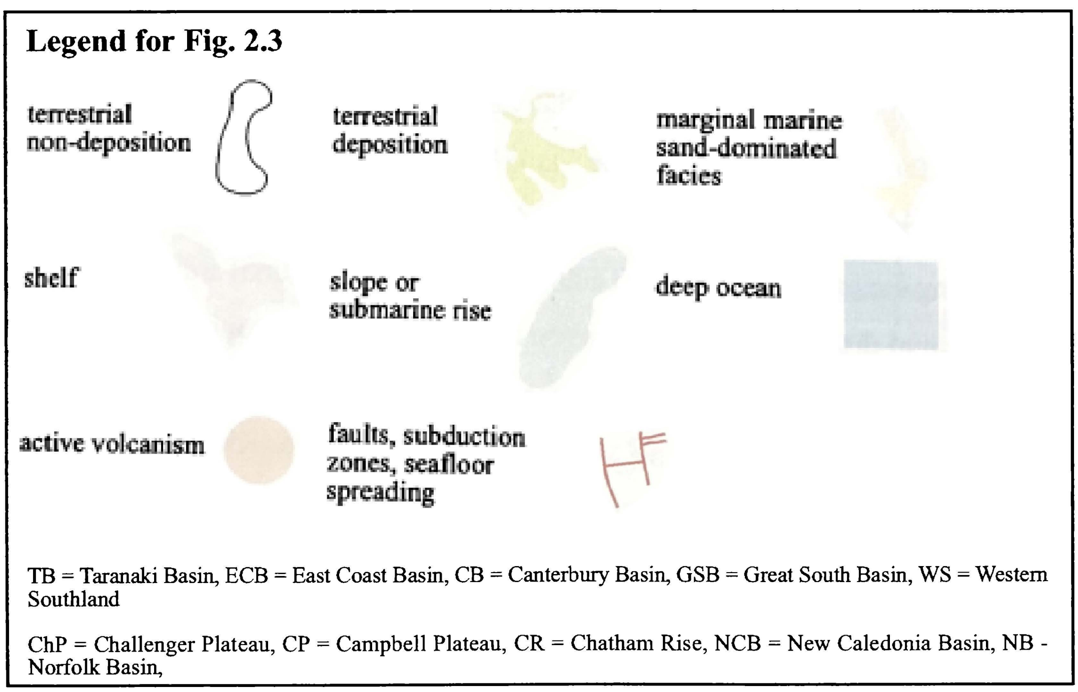
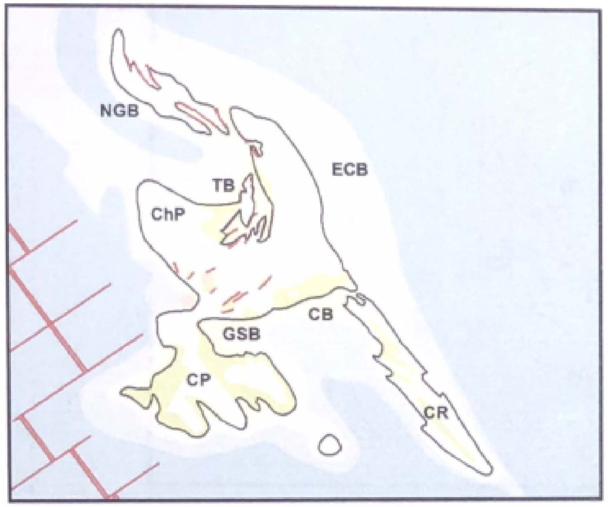
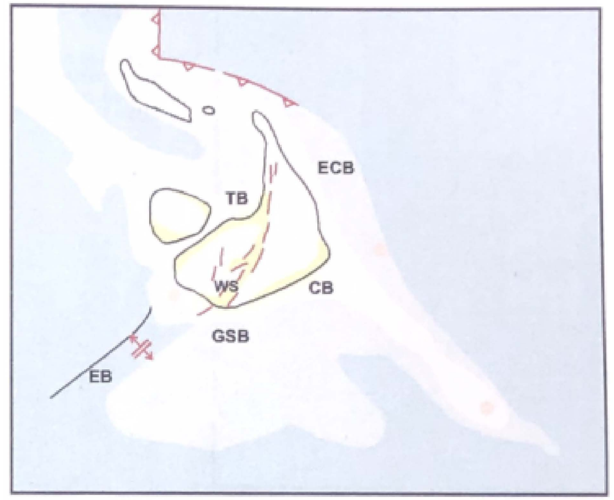


Fig. 2.3 (following pages): Paleogeographic and tectonic reconstructions of New Zealand (latest Cretaceous and Cenozoic). From King (2000a). Refer to King (2000a) for detailed text.



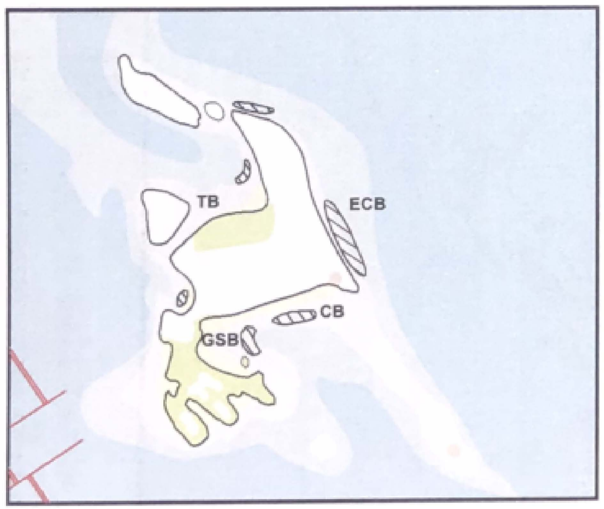
Latest Cretaceous
c.65 Ma

Syn-rift phase



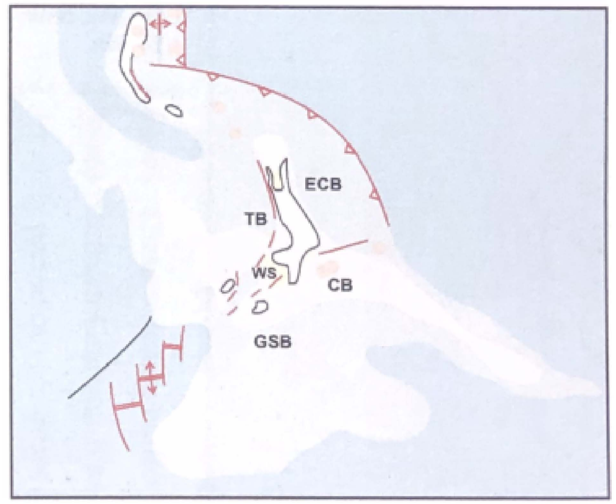
Middle Eocene
c.40 Ma

Incipient subduction north of NZ, rift propagation through western Southland to Taranaki



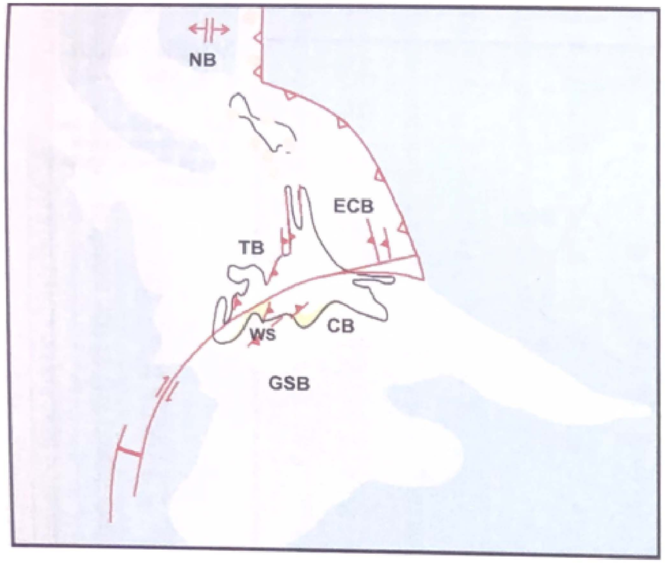
Latest Paleocene
c.55 Ma

Relatively quiescent, marine transgression of passive margin



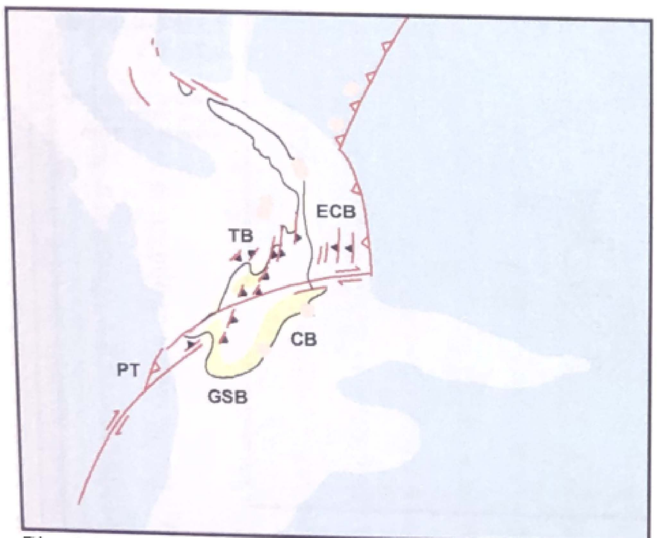
Late Oligocene
c.25 Ma

Maximun marine inundation, widespread carbonate deposition



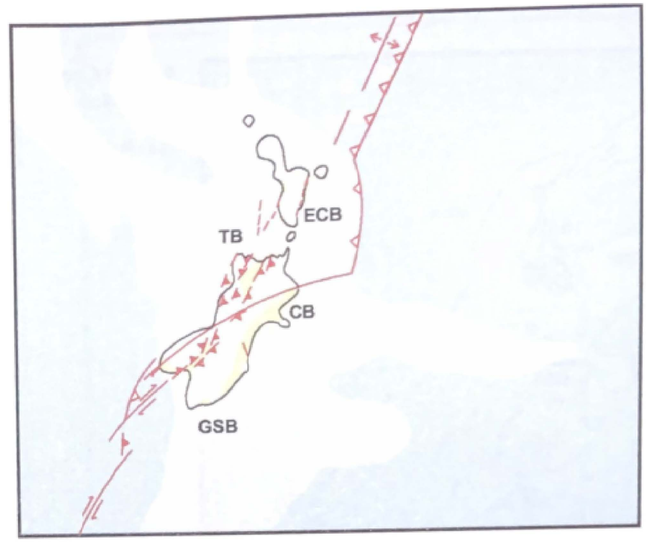
Early Miocene c.20 Ma

Propagation of Alpine Fault, North-land volcanism, compression (central NZ), fold-thrust development



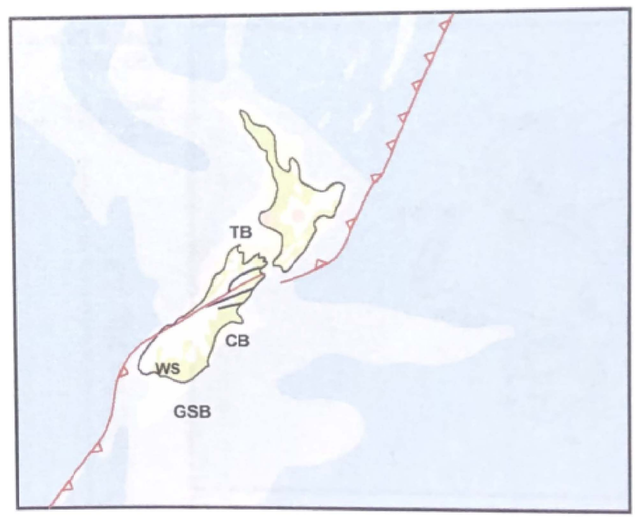
Late Miocene c.10 Ma

Arc magmatism, continued compression



Base Pliocene c.5 Ma

Extension in proto North Island, compression in proto South Island



Present Day

Subduction along Hiku-rangi and Puysegur margins. Extension in north, compression in south

Fig. 2.3 continued

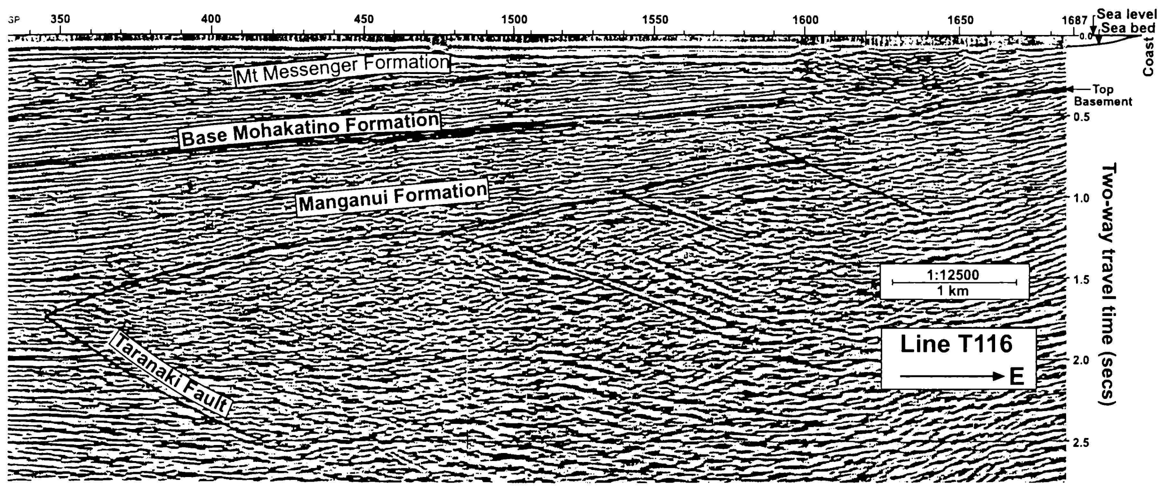


Fig. 2.4: East-west seismic reflection line T116, illustrating the thrust character of the Taranaki Fault, and onlap of Neogene sediments on to the Murihiku Terrane basement block. From King et al. (1993).

Several factors influenced Miocene sedimentation, including episodic production of volcaniclastic material from the Miocene submarine volcanoes. These volcanic edifices form a near-linear trend through the centre of the (as then unformed) Northern Graben, in an arc that lies parallel to the modern Taupo Volcanic Zone (TVZ). An early period of volcanism occurred at around 20 Ma (Kora volcanic centre), but most activity occurred from the Middle to Late Miocene (14-10 Ma), migrating from north to south (Bergman et al., 1992; Stagpoole, 1997). Volcanism is considered to have diminished significantly or ceased by about 8-6 Ma (King, 2000b). More than 25 volcanic complexes have been mapped offshore (Bergman et al., 1990), with at least 10 mapped in detail in the northern Taranaki Basin, and 6 in the Northern Graben (G. Thrasher, pers. comm., 2001; Thrasher et al., 2002; Fig. 2.5). It is estimated that volcanic edifices occupy about 15-20% of the area of the Northern Graben (Bergman et al., 1992; King and Thrasher, 1996). Typically cone or domed shaped on seismic reflection profiles, reminiscent of their andesitic composition (e.g., Kora volcanic complex; ARCO Petroleum NZ Inc., 1985, 1988), these volcanoes are thought to have erupted in a subaqueous environment, although some volcanoes display features that are characteristic of subaerial exposure, such as having bevelled tops (Chapter 5).

Volcanic massifs are progressively onlapped by Late Miocene to Early Pliocene bathyal deposits and were completely buried by the Plio-Pleistocene deposits of the Giant Foresets Formation. Paleo-topographic highs created by these edifices (estimated to range up to 1 km in thickness; King and Thrasher, 1996) had a significant impact on the distribution of sediments during the Early Pliocene, deflecting sediment pathways, as well as creating slopes down which sediment was reworked and re-deposited. To the east, Miocene sediments onlapped the lower flanks of the Herangi High.

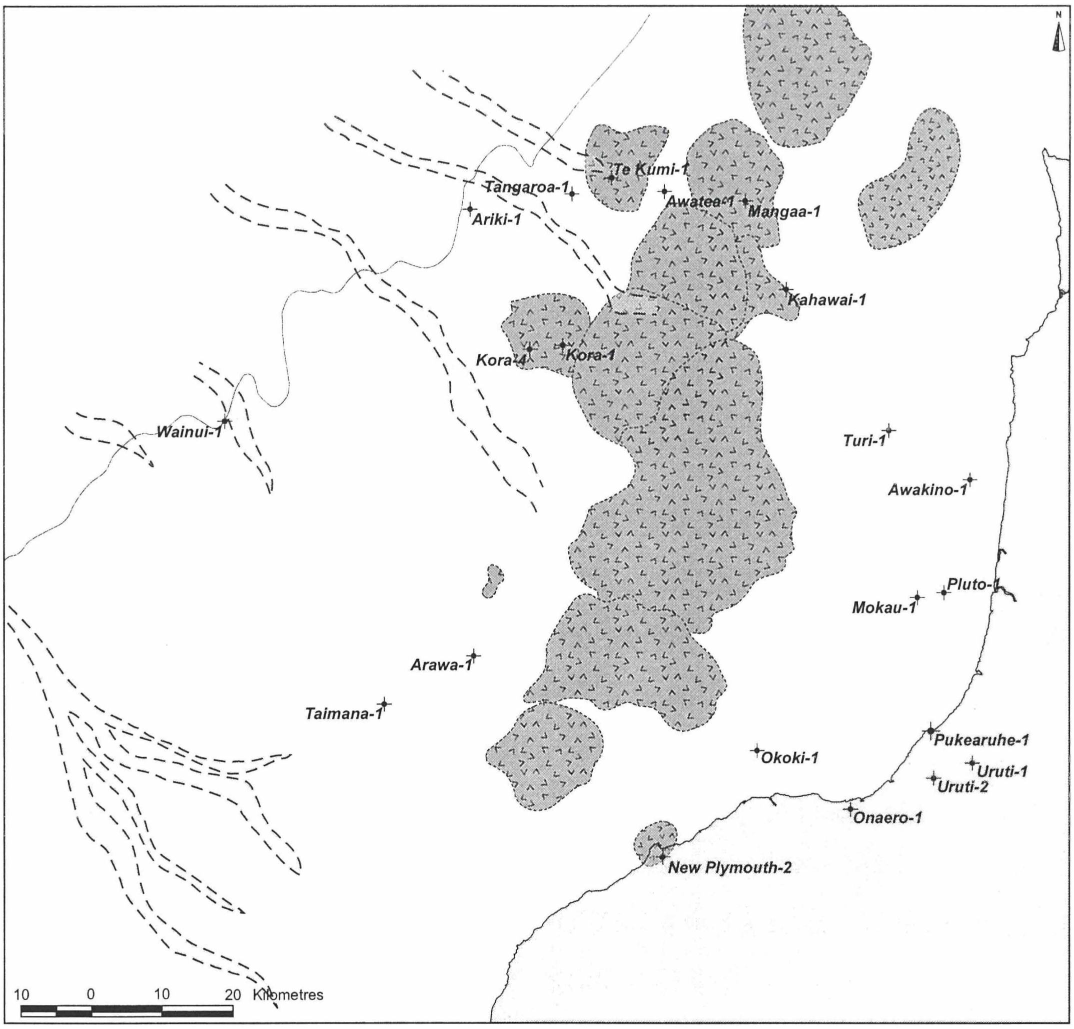


Fig. 2.5: Miocene and Pliocene volcanic centres (dark grey) and basal Pliocene channels (pale grey) mapped from seismic reflection profiles, northern Taranaki Basin. Adapted from Thrasher and Cahill (1990) and Thrasher et al. (2002).

Other influences on sediment control include the proximity and width of the paleoshelf. King and Thrasher (1996) suggest that throughout most of the Miocene, a very narrow shelf existed along the basin's eastern margin, located along the Herangi High, and opening out to the south. More recent work (Kamp et al., 2002; in prep.) suggests that during the latest Miocene and Early Pliocene the shelf was a lot more extensive in the King Country Basin and across eastern parts of Taranaki Peninsula. Underlying the shelf deposits and extending into Taranaki Basin, there were numerous submarine fan systems, with several sources inferred along the paleoshelf break. In places, these fans were also fed from volcaniclastic aprons surrounding the volcanic edifices in the centre of the basin (Nodder, 1987). In the latest Miocene, and into the Pliocene and Pleistocene, the shelf continually broadened as a result of progradation of the continental wedge. The sediments were sourced from marked uplift and erosion of the Southern Alps from about c. 5 Ma (Kamp et al., 1989; King, 2000a).

2.3.2(ii) Pliocene-Pleistocene (c. 5.2 Ma – Present)

Continuing plate-boundary deformation during the Pliocene and Pleistocene created significant areas of back-arc extension in Taranaki Basin, including the Northern and Central Grabens, as well as regions of crustal downwarp (Toru Trough and Wanganui Basin). Eastern areas underwent regional uplift and southwestwards tilting (Stagpoole, 1997). Conversely, while eastern and southern areas of the basin were undergoing active deformation, the Western Stable Platform remained relatively quiescent, with net subsidence due to lithospheric loading (e.g., Hayward, 1987; Hayward and Wood, 1989; Holt and Stern, 1991).

In the northern part of Taranaki Basin, Late Miocene to Pliocene faulting in the Cape Egmont Fault Zone and the Turi Fault Zone defined extension of the Northern Graben. The Cape Egmont Fault Zone, previously a zone of reverse faulting, was reactivated in the Late Miocene as a series of east-dipping normal faults (Stagpoole, 1997). Reactivation of this zone was primarily responsible for development of the Central Graben (King and Thrasher, 1996). The Late Miocene to Pliocene extension has been related to clockwise rotation of the Hikurangi margin, relative to a fixed Western Stable Platform, driven by migration and steepening of the subducted slab along the northern part of the Hikurangi margin (King and Thrasher, 1996).

While faulting was most active during the Late Miocene and Pliocene, present-day movement on major faults is evidenced by the offset of sediments at the sea floor (Nodder, 1993). Seismic mapping reveals that in the northern part of the Turi Fault Zone, Late Miocene sediments are offset by faults that do not offset Pliocene sediments. Conversely, to the south, Pliocene sediments are offset by faults. This indicates that the zone of active normal faulting migrated southward through the basin during the Late Miocene and Pliocene (Stagpoole, 1997), concomitant with the Plio-Pleistocene evolution of the Northern Graben. Calculated amounts of east-west Neogene extension north of the Taranaki Peninsula have been estimated at between one and three kilometres (King and Thrasher, 1996; Stagpoole, 1997). East of the Turi Fault Zone, tilting, uplift and erosion resulted in the formation of a broad submerged platform (Manganui Platform; King and Thrasher, 1996) on which a thin veneer of unconsolidated Pleistocene sediments unconformably overlies Miocene strata. At Turi-1, up to 1300 m of erosion is estimated to have taken place during the Late Miocene to Pleistocene (King and Thrasher, 1996).

The Plio-Pleistocene was characterised by extremely high sedimentation rates in Taranaki Basin (over 2000 m in 5 m.y.), with sediment predominantly sourced from erosion of the Southern Alps and the rising North Island hinterland (Beggs, 1990; King and Thrasher, 1996; Kamp et al., 2002). The continental shelf continued to broaden and advance northward, although Early Pliocene subsidence in Wanganui Basin initially controlled the rate of advancement of the progradational front (Beggs, 1990; Kamp et al., 2002). Infilling and overtopping of the Wanganui Basin depocentre resulted in accelerated northwestward progradation of the continental shelf across the Western Stable Platform and the Northern Graben during the Late Pliocene through Pleistocene, reaching the northwestern parts of the basin during the last 2 m.y. (King and Thrasher, 1996). It is this Plio-Pleistocene succession, mapped over much of Taranaki Basin (Thrasher and Cahill, 1990), that constitutes the Giant Foresets Formation.

Sediment supplied to the advancing foreset front was fed by several canyons and channels that reached upslope into shelf catchment areas (King and Thrasher, 1996). Slope canyons of Miocene to Recent age are clearly evident on seismic reflection profiles, and Thrasher and Cahill (1990) have mapped an impressive tributary network of basal Pliocene canyons and channels that presumably funnelled sediment from the shelf edge on to the New Caledonia Basin abyssal plain (Fig. 2.5). In outcrop, analogous channels have been mapped in the Late Miocene Urenui and Matemateaonga Formations (see below).

2.4 Nomenclature and lithostratigraphy

This section presents a brief review of the Neogene sedimentary fill, primarily, the Miocene Wai-iti Group, and the Plio-Pleistocene Rotokare Group. For a more complete description of lithostratigraphic units, refer to King et al., (1993), Hansen (1996), and King and Thrasher (1996).

2.4.1 Current nomenclature

King and Thrasher (1996) completed a review of the stratigraphy and nomenclature of the Cenozoic Taranaki Basin fill in their monograph (refer to their Appendix 3). Revised stratigraphies for several wells are given in Appendix 1. This includes wells for which no stratigraphy was given in King and Thrasher (1996).

2.4.2 Neogene Stratigraphic units

The Neogene stratigraphy of Taranaki Basin is summarised in Fig. 2.6. The stratigraphic units that belong to the Wai-iti and Rotokare Groups are of particular interest to this study:

1. the Miocene Wai-iti Group incorporates the Manganui, Moki (not discussed), Mohakatino, Mount Messenger, Urenui, and Ariki Formations;
2. the Plio-Pleistocene Rotokare Group includes the Matemateaonga and Tangahoe Formations, Mangaa Formation, and Giant Foresets Formations, as well as undifferentiated shelf deposits (not discussed).

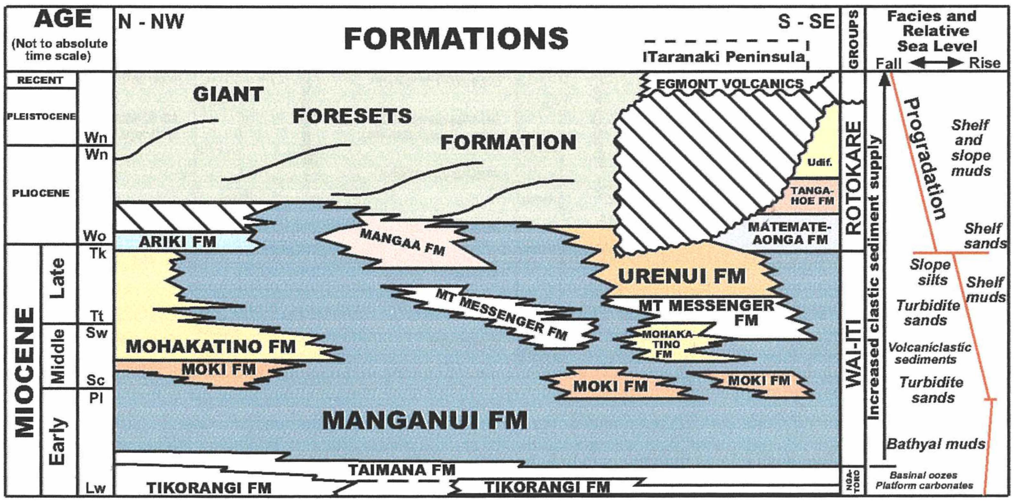


Fig. 2.6: Miocene to Recent stratigraphic framework for Taranaki Basin. This figure illustrates the general age and progradational nature of the Giant Foresets Formation, and its relationship to underlying or time equivalent formations. Modified from King and Thrasher (1996).

Kamp et al. (2002) have erected two new groups (2^{nd} order megasequences) for Wanganui Basin and King Country Basin strata: the Whangamomona Group and the Rangitikei Supergroup. The Whangamomona Group includes the Mangarara Formation, Otunui Formation, Mount Messenger Formation, Urenui Formation, Kiore Formation (new – refer to Vonk et al., 2002 for description) and the Matemateaonga Formation (Fig. 2.7). The Rangitikei Supergroup includes the Tangahoe Mudstone and various younger units that occur within the Wanganui Basin. The Matemateaonga Formation is included within the older Whangamomona

Group on the basis of a major flooding surface between this formation and the overlying Tangahoe Mudstone (see section 2.4.3 and discussion in Kamp et al., 2002). Rangitikei Supergroup units form the lateral equivalents to the Giant Foresets Formation. The relationship between the nomenclature used by King and Thrasher (1996) and Kamp et al. (2002) is summarised in Fig. 2.7.

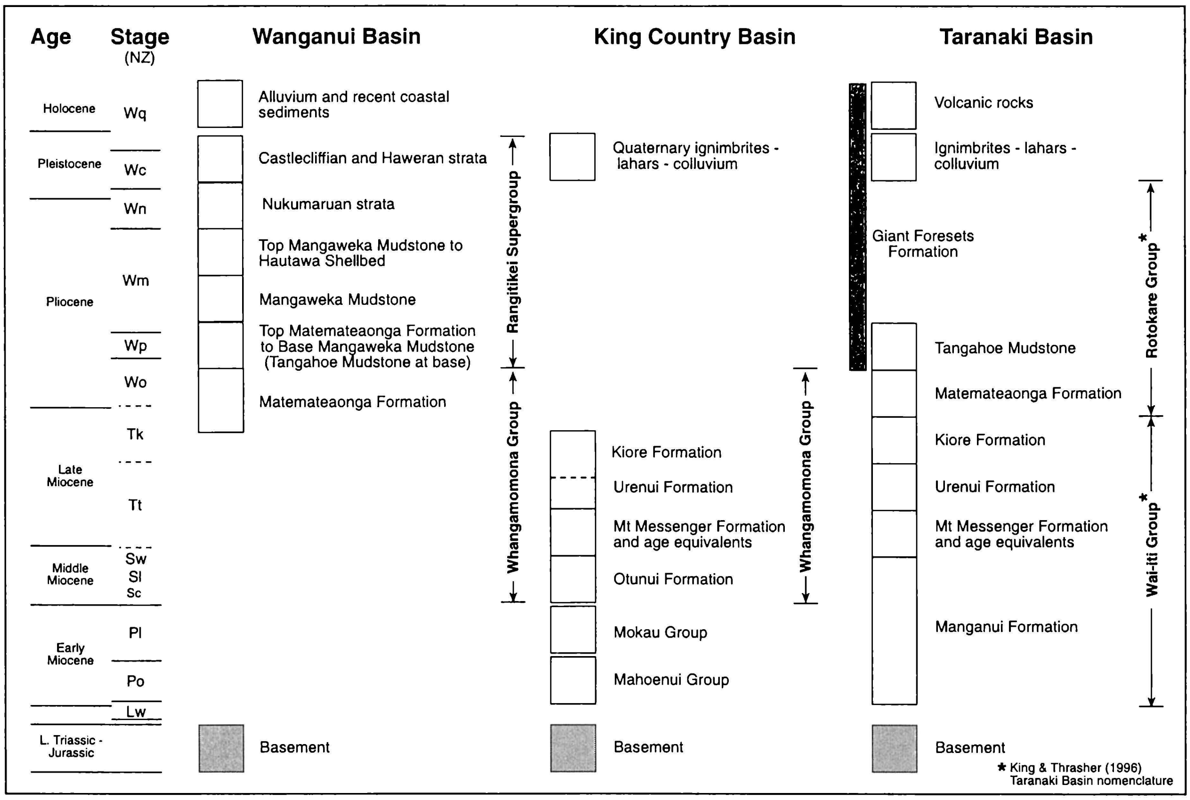


Fig. 2.7: Major stratigraphic units of Taranaki, King Country and Wanganui Basins. Note the placement of the Matemateaonga Formation in the Whangamomona Group (Kamp et al., 2002), rather than the Rotokare Group (King and Thrasher, 1996). Figure from Kamp et al. (2002).

2.4.2(i) Miocene Wai-iti Group

The marine-dominated Wai-iti Group was introduced by King (1988a,b) to incorporate most of the Miocene sedimentary succession in the Taranaki Basin. The name is derived from Wai-iti Beach along the northern Taranaki coastline. It is along this coastline that constituents of the group (Mohakatino, Mount Messenger, and Urenui Formations) are spectacularly displayed in outcrop. The Wai-iti Group represents deposits of a regressive Miocene megacycle. It is clastic dominated, and includes shelf, slope and basin floor mudstone and siltstone (Manganui Formation), sandstone-dominated redeposited beds (Moki and Mount Messenger Formations), slope siltstone (Urenui Formation), volcaniclastic dominated successions (Mohakatino Formation), and marly siltstone of the Ariki Formation (King and Thrasher, 1996).

Wai-iti Group sediments, particularly the Mount Messenger Formation, cropping out along the northern Taranaki coastline are important as they provide analogues for subcrop strata, and allow depositional mechanisms, paleoenvironments, lithologies, and the internal sequence architecture to be determined.

Manganui Formation

Initially introduced by King (1988a,b) to incorporate deep-water mudstone beds that dominate the Miocene succession in Taranaki Basin, definition of the Manganui Formation has since been expanded to include all fine-grained rocks that do not fit into any other formation (King and Thrasher, 1996). The Manganui Formation is therefore correlative with parts of the Mount Messenger, Urenui, and Giant Foresets Formations. Subdivision of the Manganui Formation is sometimes convenient, and the term 'lower' Manganui Formation was adopted for the interval between the Taimana and Moki Formations. Where the Manganui Formation becomes indistinguishable from the Giant Foresets Formation, the top of the Manganui Formation is arbitrarily taken as the top Miocene seismic horizon, as calibrated by biostratigraphic methods in well sections.

The introduction of muddy sediment comprising the Manganui Formation led to the start of regressive sedimentation in Taranaki Basin. However, despite the accumulation of Manganui Formation mudstone, bathyal depths persisted across most of Taranaki Basin during the Early Miocene to middle-Late Miocene. The Manganui Formation is voluminous and widespread, and within some wells in the study area, reaches a thickness greater than 1600 m (e.g., Taimana-1). At many sites in the northern part of Taranaki Basin, it interfingers with volcaniclastic sediments of the Mohakatino Formation, and at these sites the composite name of (upper) Manganui-Mohakatino Formation is used.

Mohakatino Formation

Hay (1967) erected the Mohakatino Group (later emended to Mohakatino Formation in King et al., 1993) for Middle to Late Miocene strata characterised by a significant tuffaceous or volcaniclastic content. Onshore, these were subdivided into four formations: Mangarara Sandstone, Purupuru Tuff, Ferry Sandstone and Tawariki Mudstone. Later studies placed the Ferry Sandstone and Tawariki Mudstone in the overlying Mount Messenger Formation (see Hansen, 1996; King and Thrasher, 1996). The volcaniclastic component within middle-Late Miocene strata diminishes dramatically to the east, away from source, with the result that the stratigraphic subdivisions of Hay (1967) are not identifiable in age equivalent strata only a few kilometres inland (King et al., 1993). Offshore, in the northern part of Taranaki Basin, volcaniclastic sediments are more widespread, distributed primarily throughout the northern and northeastern parts of the basin, proximal to the Mohakatino Volcanic Centre. Generally volcaniclastic sediments are clearly evident on wireline log responses, and where present are assigned to the Mohakatino Formation. Occasionally, volcaniclastic/non-volcaniclastic lithotypes are indistinguishable, and in some well sections sediments are assigned to undifferentiated Mohakatino/(upper) Manganui Formation. Both in outcrop and subcrop, the volcaniclastic content diminishes upsection, partly reflecting a waning of volcanic activity (Shell BP Todd Oil Services Ltd., 1984), and partly masked due to the rates of accumulation of non-volcaniclastic detritus.

An extended period of eruptive activity is evident from the range of ages determined for the Mohakatino Formation. This is estimated to have ranged from Early to Late Miocene, with the most abundant volcanism from 21-15 Ma (Bergman et al., 1990, 1992). Eruptions from the Mohakatino volcanic centres progressed in a roughly north to south fashion, with a commensurate younging of the age of the formation to the south. Most eruptions are considered to have been of a subaqueous nature, although there may have been the possibility of subaerial activity as well. In addition to primary eruptive products, marine and airfall volcaniclastic sediments were also reworked, possibly as turbidity currents down-slope of the volcanic edifice (e.g., Nodder, 1987).

While the Mohakatino Formation is defined by the abundance of volcanic ejecta, mainly of a basaltic-andesitic composition (Bergman et al., 1990), deposition of volcaniclastic detritus was superimposed on a background of siliciclastic sedimentation. The most highly volcaniclastic successions generally occur at the base of the formation (both in offshore wells, and coastal outcrop). In the coastal section, Mohakatino sediments derived from the north at first interfingered with, and were then overwhelmed by, siliciclastic sediment prograding into the region from the south and east (Mount Messenger Formation). Well reports suggest that interfingering of volcaniclastic and non-volcaniclastic sediments occurs in the offshore region as well. Thin volcaniclastic horizons (tuffs) are interspersed throughout the overlying Mount Messenger and Urenui Formations, indicating that there were intermittent volcanic eruptions at least until the end of the Tongaporutuan (King et al., 1993).

Mount Messenger Formation

Hay (1967) proposed the name Mount Messenger Sandstone to replace the Waikiekie Formation of Glennie (1958) to prevent confusion with the Waikiekie Tuffaceous Sandstone of Jurassic age in the vicinity of Kawhia Harbour. Nevertheless, the older name prevailed in oil company literature for some time, inherited from reconnaissance work and drilling in the King Country and Taranaki Basins in the 1950s and 1960s. King (1988a,b) later emended the name to Mount Messenger Formation. The definition of the formation is generally extended to encompass all well-developed, deep-water sandstone successions of late Lillburnian to Tongaporutuan age in the basin. It is best developed in the onshore part of the basin, measuring about 850 m thickness in total (King and Thrasher, 1996) and is exposed along the north Taranaki coast between Pariokariwa Point and Mokau River, and areas immediately inland (Fig. 1.2). Quaternary tilting and erosion limit the present-day northern extent of exposure of the formation, although thick-bedded sandstone intervals, which display an affinity to Mount Messenger Formation sediments, interfinger with Mohakatino Formation successions north of the Mokau River (King et al., 1993).

In the coastal section, the lower part of the Mount Messenger Formation is characterised by thick-bedded and amalgamated sandstone beds, overlain by thin-bedded sandstone packages, generally totalling a few tens of metres thickness, and topped by a mudstone interval several metres thick. These stacked fining-upwards cycles are attributed to basin floor fan development and abandonment at the base of the advancing slope (King et al., 1993, 1994). Detailed interpretation of wireline log facies in Urenui-1 (King et al., 1993) indicates that individual sandstone and stacked sandstone bodies are fan lobe sheet or fan channel deposits, with thinner-bedded units representing deposition on the fan lobe fringe (King et al., 1994). In contrast, the upper interval is characterised by strongly interbedded, normally graded units that exhibit some or all of the classic Bouma turbidite divisions. These repeated, smaller-scale, coarsening upwards sub-cycles reflect crevasse splay or levee overbank deposition away from the main sandstone depocentres or transport axes at the base of the advancing slope (King and Thrasher, 1996). This interval is associated with interleaved and discontinuous clastic-filled channels, often with a channelised conglomerate at their base. The entire Mount Messenger Formation succession is considered to represent the lowstand deposits of 4^{th}-order (400 ka) cyclicity (King et al., 1994).

Urenui Formation

Glennie (1958) established the type section for the Urenui Formation near the township of Urenui in north Taranaki. While Hay (1967) emended the name to Urenui Siltstone, King (1988a,b) preferred the former name, and his nomenclature is followed in this study. In northeastern onshore regions and in the subsurface of the Taranaki Peninsula, the Urenui Formation forms a distinctive unit of characteristically fine-grained slope deposits that stratigraphically overlie deep-water Mount Messenger Formation sandstone beds, and underlie shelf sandstone of the latest Miocene - Pliocene Matemateaonga Formation. The Urenui Formation is contiguous with, and a lateral correlative of, upper Manganui Formation slope deposits identified in central and southwestern parts of the basin from seismic reflection and paleobathymetric signatures (Fig. 2.6). The formation is restricted to the northern Taranaki coastal strip, offshore areas immediately to the west, and to the Taranaki Peninsula region (King and Thrasher, 1996).

Channel networks within the Urenui Formation are evident both in outcrop and on nearby offshore seismic reflection profiles. In outcrop, channel systems are generally between 10-30 m thick, are characterised by deeply incised bases, and varying coarse-grained lithologies, including thick-bedded and amalgamated sandstone, thin-bedded, wavy-laminated, scoured and carbonaceous sandstone, and debris flow conglomerate (King et al., 1993). In seismic reflection profiles, subsurface channels are 1-2 km across, and many are asymmetric in cross-section (King and Thrasher, 1996). They represent sediment pathways, which have been interpreted by King et al. (1993) as infilled slope canyon complexes that transported fluvial- and shelf-derived sediments to the basin floor.

Ariki Formation

The Ariki Formation was first introduced by King (1988a,b) to describe a condensed interval of highly calcareous marl encountered in several northwestern wells (Ariki-1, Tangaroa-1, Te Kumi-1, Wainui-1 and Tane-1). It represents a significant period of terrigenous sediment starvation, attributed to a depositional hiatus caused by elevation of the Western Stable Platform relative to the actively subsiding Northern Graben (King and Thrasher, 1996), and thus delineates a paraconformity between Miocene and Pliocene sediments. While comparatively thin (attaining a maximum thickness of only 109 m), the Ariki Formation forms a bold and correlatable marker horizon on seismic reflection profiles, and is often used for mapping the base Pliocene horizon (yellow reflector of Thrasher and Cahill, 1990).

The Ariki Formation has been previously correlated with the Urenui Formation (e.g., Tangaroa-1; Shell BP and Todd Oil Services Ltd., 1981) because of its presumed age, its stratigraphic position (directly above the Mohakatino Formation), and its perceived lithological similarity (from wireline response) with the Urenui Formation in onshore north Taranaki. However, given that the Urenui Formation cannot realistically be traced far from its type area, and that the Ariki Formation is both younger in age and more calcareous than the Urenui Formation, deep-water calcareous sediments in offshore regions once attributed to the Urenui Formation are now incorporated into the Ariki Formation.

As discussed in later chapters, the Ariki Formation is distinguished by more than its calcareous and marly nature. It displays a distinctive wireline character (Chapter 4), forms a bold seismic reflector horizon (Chapter 5) and is associated with a dramatic and sudden increase in planktonic foraminiferal content downhole relative to the overlying Giant Foresets Formation (Chapter 6). Both wireline characteristics, and the increase in planktics, can be used to infer Ariki Formation age-equivalents, or other marly units, where no such unit is noted on well completion reports. Recognition of characteristic wireline log motifs (see Chapter 4) suggests that calcareous units may be more common than previously thought.

2.4.1(ii) Pliocene-Pleistocene Rotokare Group

The Rotokare Group was introduced by King (1988a,b) to incorporate the mostly Pliocene-Pleistocene aged sedimentary succession in Taranaki Basin. The distinction between the Rotokare and Wai-iti Groups is partly chronostratigraphic, but is also clearly lithostratigraphic in many places. In the southeast of the basin, the boundary is demarcated by an erosional unconformity, while to the northwest of the basin, it is associated with the Ariki Formation.

Within the Northern Graben, latest Miocene to Early/mid Pliocene strata are absent across volcanic edifices, which were areas of positive relief on the sea floor at the time. Central parts of the basin (around Taimana-1, Arawa-1) were areas of ongoing slope progradation and prolonged clastic sedimentation and here the Pliocene and Miocene successions are (relatively) contiguous and conformable. In these areas, the base Pliocene time-line (tied to well biostratigraphy) is used to separate the Miocene and Pliocene intervals. In the northeast, volcaniclastic sediments, deposited over a wide area through to the end of the Miocene, are absent in the Pliocene, providing another lithological distinction between the Rotokare and Wai-iti groups.

Plio-Pleistocene successions are thick everywhere except in the southern Taranaki Basin. Beneath the South Taranaki Bight and the southern Taranaki Peninsula, the succession is up to 2000 m thick, and east of the buried Patea-Tongaporutu High increases to 4000 m. Plio-Pleistocene accumulations up to 3000 m thick are present in the Central and Northern Grabens, and up to 2200 m thick over the Western Stable Platform (Crundwell et al., 1994; King and Thrasher, 1996).

Mangaa Formation

The Mangaa Formation, after King and Thrasher (1996), derives its name from latest Miocene to Pliocene-aged sandstone encountered at Mangaa-1, in the Northern Graben. The deep-water Mangaa Formation sandstone is younger and geographically separated from sandstone of the Mount Messenger and Moki Formations. They constitute the main body of initial sedimentary fill (submarine fan deposition) within this part of the Northern Graben during the Early Pliocene (King and Thrasher, 1996). The only other well section in which Mangaa Formation sandstone is encountered is in neighbouring Awatea-1, although there is a suggestion that well-developed, deep-water sandstone of Late Kapitean to Opoitian age in Arawa-1 may be correlatives of the formation in Mangaa-1 (Forder and Sissons, 1992; G. Thrasher, pers. comm., 2002). However, these could also be part of an intermediate depositional system, both in age and location, between the main Mangaa Formation and Mount Messenger Formation sandstone depocentres. Thinner, less well-developed sandstone of Opoitian age also occur nearby in Taimana-1 (Forder and Sissons, 1992). At both Mangaa-1 and Awatea-1, the Mangaa Formation unconformably overlies the Mohakatino Formation, and underlies the Giant Foresets Formation.

Forder and Sissons (1992) divided the Mangaa Formation at Mangaa-1 into two intervals: Moki C1 Sands (youngest) and Moki C2 Sands (oldest). Together the sandstone-dominated interval is greater than 750 m thick, with each interval separated by a thick mudstone interval nearly 100 m thick. Forder and Sissons (1992) also recognised a third older and petrographically distinct interval that has since been incorporated into the Mohakatino Formation (see King and Thrasher (1996), their Appendix 3, Mangaa-1). At Awatea-1, three sand intervals were also identified (Mangaa A, B, and C sands; Murray and de Bock, 1996). The lowest Mangaa C sands are Tongaporutuan in age – the age and stratigraphic position suggest that they could correlate on an age basis with the Mount Messenger Formation. However, because the Mount Messenger Formation is not know that far north, Mangaa C sands are included in the Manganui Formation, (intra-Manganui sandstone; see Appendix 1, Awatea-1). Mapping of the Mangaa Formation using seismic reflection profiles (Forder and Sissons, 1992) has revealed a wedge-shaped geometry that thins to the northeast, and is thickest along the axis of the Northern Graben.

Giant Foresets Formation

The Giant Foresets Formation was first given formation status by Shell, BP and Todd (1977; in Pilaar and Wakefield, 1978). They described a striking series of giant clinoform-shaped, basinward-dipping foresets beds, with corresponding topset and bottomset facies (Beggs, 1990) imaged on seismic reflection profiles (Fig. 2.8). The name has been widely used, but no formal definition of the formation has been published. In part, this is due to the name itself: theoretically, 'foresets' imply only the foresetted part of the succession, however, the informal definition includes both topset and bottomset facies. To formalise a name for the deposits, it would be necessary to either (a) devise a new name that covers *all* the relevant Plio-Pleistocene sediments deposited as part of the prograding continental wedge, or (b) split the formation into three distinct parts, each given separate formational status. The first option, while probably the easier of the two to achieve, means discarding a formational name that has been commonly used in the past and would result in considerable confusion. The second option is difficult because of the diachronous nature of the succession. Formational boundaries would thus be considerably different in age and stratigraphic position from site to site, depending on the timing of progradation through that site. King (1988b) suggested retaining an informal formation status for the Giant Foresets Formation, which is followed in this study.

The Giant Foresets Formation represents ongoing aggradation and progradation of the post Miocene continental shelf wedge, although a simplified model of progressive northwards outbuilding and widening of the continental shelf is complicated by the late Neogene extension and graben formation that affected northern Taranaki Basin. In particular, subsidence of the Northern and Central Grabens initially funnelled and trapped sediment in these early depocentres, with layer-cake infill and aggradation (e.g., Mangaa Formation in the Northern Graben). Differential fault movements along the Turi and Cape Egmont Fault Zones also produced widely varying stratigraphic thicknesses of the Plio-Pleistocene sequences. For example, post Miocene sediments reach nearly 3000 m thickness in the depocentre to the east of Arawa-1, but may be less than 100 m thick over the Manganui Platform.

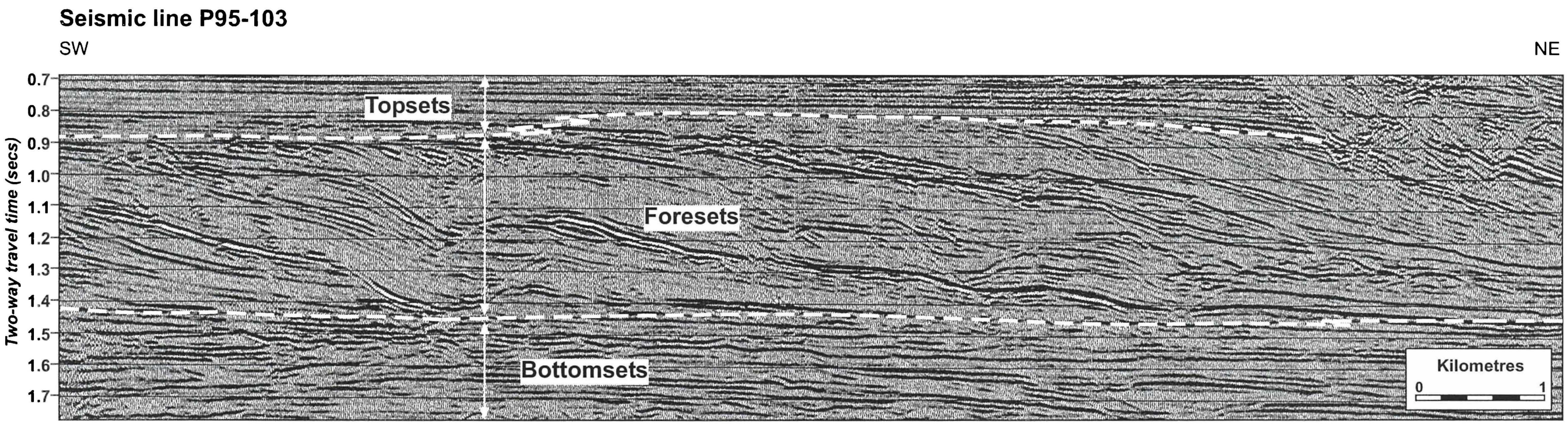


Fig. 2.8: Part of seismic reflection profile P95 103, illustrating the seismic expression of the Giant Foresets Formation. The formation, as informally defined, incorporates basin floor facies (bottomsets), slope facies (foresets) and shelf facies (topsets). Refer to Chapter 5 for a complete description of the seismic characteristics of this formation.

On seismic reflection profiles, bold, high amplitude seismic reflectors that delineate clinoforms are presumed to be horizons of partial lithification of the paleo-sea floor, representing flooding surfaces and condensed sections formed during high relative sea level (refer to Chapter 5). Each successive clinoform represents the transient position of the shelf-slope-basin floor profile, and are thought to have a similar depositional cyclicity (4^{th}-order) to comparable sequences described in the outcropping Mount Messenger Formation (Beggs, 1990; King et al., 1994). The Giant Foresets Formation cyclicity may be 5^{th} to 6^{th} order, 100 000 or 41 000 ka cycles. Shelf margin offlap was directed towards the west and northwest, and during the Plio-Pleistocene, the shelf migrated between 8-20 km in that direction (King and Thrasher, 1996). Because of the progradational nature of the formation, facies belts within the Giant Foresets Formation are diachronous, with thick Pliocene shelf accumulations in southern and central parts of the basin, and slope to basin floor deposits elsewhere. Pleistocene sediments tend to be shelfal everywhere, extending to the present day shelf edge, except for the northwestern area, where they also encompass thick slope accumulations (Beggs, 1990).

Examination of well reports suggests that a lithological distinction can be made between the upper part of the Giant Foresets Formation, and the lower part. At Ariki-1, three lithostratigraphic units are identified (Shell BP Todd Oil Services Ltd., 1984):

1. 474-950 m - thick, poorly defined beds of siltstone and silty mudstone. Muds are light greenish grey to medium grey, sticky and plastic with distinct black specks (carbonaceous material?). Siltstone units are similar in colour, though occasionally yellowish brown. Sporadic calcareous stringers and lenses occur. Foraminifera are rare to common, and occasional layers rich in macrofossils occur.
2. 950-1335 m - siltstone and silty mudstone units as above, with graded beds recognised from log character.
3. 1335-2147m – light grey silty mudstone, slightly harder and more consolidated than above, being sticky but not plastic. Foraminifera are common throughout, with only rare horizons of macrofossil debris.

Several other well reports make reference to sedimentary distinctions within the Giant Foresets Formation. At Tangaroa-1, Kora-1, and Kora-4, two informal units have been recognised; an upper sequence comprising mainly light grey, soft to firm, slightly calcareous, quartz siltstone or claystone, intercalated with minor fine argillaceous sandstone of similar colour, with scattered macro- and microfossil debris, and a lower sequence of light to dark greenish grey, soft to firm, and occasionally calcareous mudstone, with rare thin silty or sandy horizons, and trace fossil fragments (Shell BP Todd Oil Services Ltd., 1981; ARCO Petroleum NZ Inc., 1985, 1988). At Wainui-1, the interval above 700 m consists of predominantly coarse-grained, shelly sand with thin siltstone and mudstone, and below 700 m, is dominated by moderate grey mudstone with subordinate siltstone. While the stratigraphy at Arawa-1 is more diverse, probably reflecting the sites closer proximity to the paleo-shoreline, two broad intervals can also be identified. As with the previous well sections, a coarser upper interval, dominated by interbedded sandstone, siltstone, and claystone, with abundant shell fragments, and a lower interval of more massive, light grey to greenish-grey sticky claystone (ARCO Petroleum NZ Inc., 1992) can be distinguished.

Shell BP Todd Oil Services Ltd. (1984) suggest that the three lithostratigraphic units identified at Ariki-1 represent a gradual but pulsed basin evolution from deep bathyal (lowermost unit) through to upper bathyal to outer neritic (middle unit) to neritic paleoenvironments (uppermost unit). Subsequent chapters expand on this idea, investigating the link between lithology, geophysical properties, and paleoenvironmental setting.

2.4.3 Stratigraphic relationship between Taranaki and Wanganui Basins

Several units encountered in Wanganui Basin are considered to be lateral equivalents to units within the Taranaki Basin (refer to Fig. 2.7), having been part of the same depositional system. Kamp et al. (2002) recognise four 2^{nd}-order cycles in King Country and Wanganui Basins equivalent to the Early Miocene-Recent regressive megacycle of King et al. (1994). The younger two megasequences comprise the Whangamomona Group/megasequence (including Otunui, Mount Messenger, Urenui, Kiore, and Matemateaonga Formations) and the Rangitikei Supergroup/megasequence (including the Tangahoe Mudstone and undifferentiated Whenuakura Subgroup and younger units). The Whangamomona Group represents an older progradational phase that occurred *prior* to the present progradational phase that has resulted in the deposition of the Giant Foresets Formation and outbuilding of the modern continental shelf. Terrigenous starvation in the northern Taranaki Basin (represented by the condensed Ariki Formation and equivalents) is in part attributed to the limited extent of progradation of the "Whangamomona" continental margin into Taranaki Basin during this earlier phase.

Later parts of paraconformity development in Taranaki Basin are related to a marked Early Opoitian (earliest Pliocene) tectonic pulldown of Wanganui Basin and Toru Trough. This caused a dramatic flooding and southward onlap across the basement of Wanganui Basin and southeastern parts of Taranaki Basin. Water depths increased dramatically from shoreface to bathyal, causing accumulation of Matemateaonga Formation sediments to cease, and the deposition of Tangahoe Mudstone to start. Contemporary shelf sedimentation was focused to the south in Wanganui Basin, while to the north the King Country Basin was undergoing doming, probably as a lithostatic response to thermally driven uplift associated with the migration of the volcanic arc from Taranaki Basin into the Taupo Volcanic Zone (TVZ; P. Kamp, pers. comm., 2003). The Tangahoe Mudstone signifies slope migration of a progradational wedge (Rangitikei Supergroup/megasequence), this time on two fronts; one northward through Wanganui Basin and into King Country Basin, and a second west of the Patea-Tongaporutu High, moving progressively from the southern to the northern parts of Taranaki Basin (Kamp et al., 2002). Bottom-sets and slope-sets of the Tangahoe Mudstone pass upwards into shelf deposits of the Whenuakura Subgroup, and together constitute the southern and eastern forerunner to the younger Giant Foresets Formation in northern Taranaki Basin.

The nature of the relationship between Taranaki and Wanganui Basins (and also the King Country Basin) is indicated by progressively younger Miocene strata overtopping the Patea-Tongaporutu High, which during the Middle Miocene formed a narrow, elevated or shallow submarine basement ridge that separated Taranaki and Wanganui Basins (King and Thrasher, 1996). In the Taranaki Peninsula region, Late Miocene and Pliocene strata onlap and overtop basement. This demonstrates that the Taranaki Basin was contiguous with the King Country Basin (as defined by Kamp et al., in prep.) in the Late Miocene at least, and with the Wanganui Basin in the Plio-Pleistocene (King and Thrasher, 1996). Subsequent Quaternary uplift and deformation tilted the Neogene section in eastern Taranaki Basin to the southwest, and resulted in the complete erosion of Pliocene strata in the northern parts of Taranaki Peninsula.

2.5 Provenance

Few detailed provenance studies have been conducted on Neogene sediments. Most information is gleaned from Well Completion Reports, and the occasional composite report that disseminates information from original well reports (e.g., Smale, 1996). However, rare reports have included additional laboratory research (e.g., Bergman et al., 1990), while Hudson (1996) concluded that Neogene formations cropping out along the northern Taranaki coastline could be placed into one of three groups based on their mineral assemblages:

1. Calcareous to non-calcareous shelf or shelf-derived sandstone and flysch (Oligocene to Middle Miocene rocks of inland northern Taranaki) that have a predominantly local North Island greywacke source, and volcanogenic sandstone sourced from offshore andesitic volcanism.
2. Middle to Late Miocene rocks which contain a high abundance of phyllosilicate minerals, microcline and schistose rock fragments, and which suggest an increasing supply of metamorphic and granitic derived material being delivered into the northern Taranaki area (including sediments of Mount Messenger, Urenui and Matemateaonga Formations). This has been linked to increasing rates of uplift forming the Southern Alps, and their rapid erosion.
3. Pliocene sandstone that contain less metamorphic derived-material and more greywacke-derived material interpreted to be sourced from the Wanganui Basin. This is suggested to reflect uplift of southern North Island ranges, providing an increasing supply of greywacke material.

Other studies also indicate that Early Miocene sediments were sourced from uplift and erosion of the hinterland (Herangi Range; Bennett et al., 1992; King and Thrasher, 1992), with an increasing component derived from erosion of the emerging Southern Alps during the Middle to Late Miocene. Uplift of the Tararua/Ruahine ranges during the Clifdenian to Lillburnian (~ 15 Ma) is also suggested as a source for Middle Miocene sediments (e.g., inland Moki and Otunui Formations; Kamp et al., 2002). Andesitic volcanism contributed substantially to Middle to Late Miocene sediments, particularly in offshore regions (Mohakatino Formation).

As volcanism ceased and the regressive megacycle took hold, siliciclastic sedimentation sourced from the South Island dominated. In northern Taranaki Basin, this was characterised early on by the development of large basin floor fan systems (Moki and Mount Messenger Formations) followed by slope deposits (Urenui Formation) that infilled the foredeep to the west of the Taranaki Fault. Prodigious amounts of finer-grained sediment (Manganui Formation) dominated deeper-water sedimentation until the Late Miocene. Wai-iti Group sediments have a low-grade metamorphic provenance, suggestive of derivation from a Western Province (Fig. 2.9) and/or Marlborough Schist (Smale, 1996). The presence of stilphnomelane,

stilphnomelane, a mineral restricted to low-grade schists from the northwestern part of the South Island (Jordan et al., 1994) in the Mount Messenger Formation supports a southern provenance (see also Robert et al., 1986).

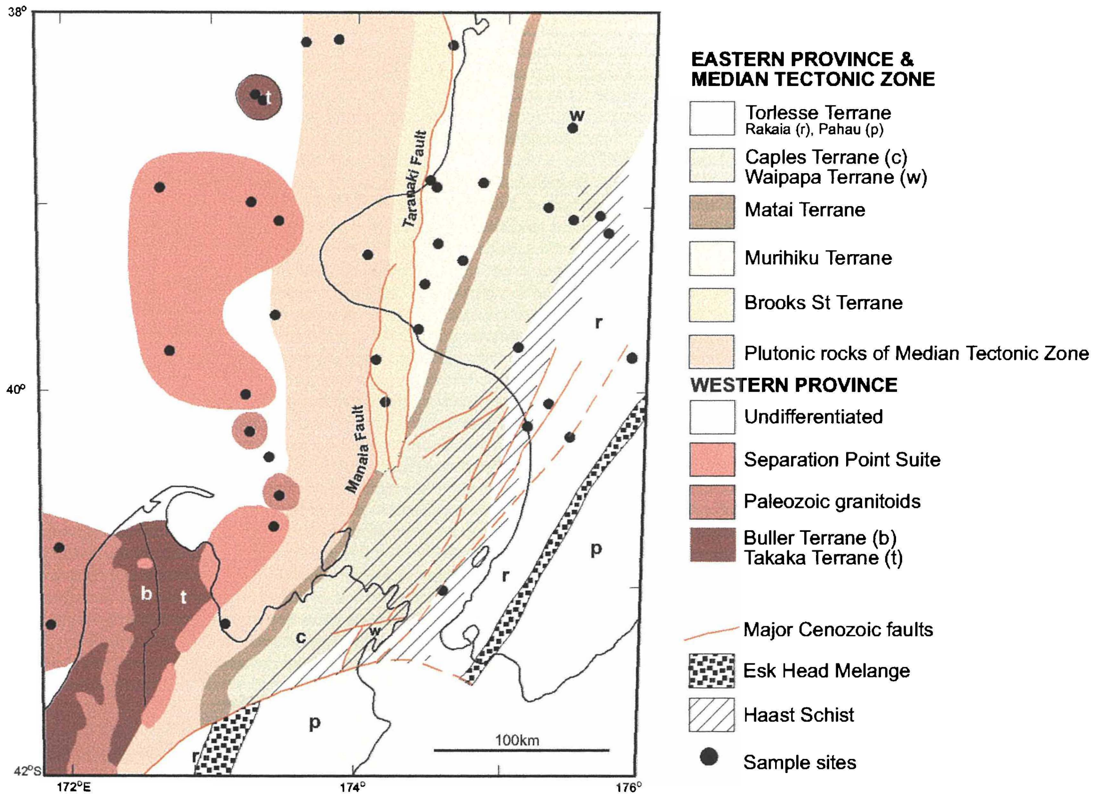


Fig. 2.9: Pre-late Cretaceous basement geology beneath Taranaki, Wanganui and King Country Basins, showing distribution of Eastern and Western Province terranes. Adapted from Mortimer et al. (1997). Refer to Mortimer et al. (1997) for description of terranes.

While little attention has been given to post Miocene sediments in petroleum reports, gross examination of the mineral assemblage of Plio-Pleistocene sediments in southeast Taranaki Basin suggests that Mesozoic strata (emergent Southern Alps) were the dominant provenance for this succession (Forder and Sissons, 1992). This concurs with Beggs (1990), who concluded that most of the Plio-Pleistocene sediment deposited in the Giant Foresets Formation was derived from the south, primarily from the rising Southern Alps. This conclusion was based on the overall offlap direction, variations in continental margin morphology, facies distribution patterns, and by analogy with modern sediment dispersal systems driven by longshore drift, rather than on a lithological basis. However, Forder and Sissons (1992) also suggest that sandstone beds of the Early Pliocene Mangaa Formation, which underlies the Giant Foresets Formation in the Northern Graben, may have been partially sourced from erosion of Murihiku

Supergroup strata (i.e., east of Taranaki Fault; Fig. 2.9). It is also suggested here that the South Island signature characterising Mangaa Formation sandstone is acquired from erosion of Whangamomona Group sediments in King Country Basin, rather than directly from a South Island provenance.

Chapter 3: Biostratigraphic review of northern Taranaki Basin

Chapter 3: Biostratigraphic review of northern Taranaki Basin

3.1 Introduction

Foraminiferal sampling was primarily undertaken for reconstruction of the paleo-environmental history of the Giant Foresets Formation (Chapter 6). In addition, the opportunity to review the Late Tongaporutuan to Castlecliffian biostratigraphy of four well sections was taken. Given the limitations of using unwashed well cuttings compared with core samples (discussed in section 3.3), the two main objectives for this chapter are to, (1) refine or constrain the stratigraphic position of the New Zealand stage boundaries in the well sections; and (2) to integrate these revised stage boundary positions with biostratigraphic information from other well sections in the region to construct a robust chronostratigraphic framework for the area. As illustrated in Fig. 2.6 and discussed in Chapter 2, the base of the Giant Foresets Formation is markedly diachronous, being older to the south and younger to the north, consistent with northwestwards progradation. To the south, the base of the formation appears to be continuous with underlying formations, but to the north, its base is usually marked by a Miocene-Pliocene unconformity, which is associated with the highly condensed and seismically prominent marly Ariki Formation, or its equivalents. Thus, a third objective of this chapter, through foraminiferal analysis, is to identify and gain an appreciation of the nature and significance of the contact between the Giant Foresets Formation in northern Taranaki Basin and the formations underlying it.

The selection of well sections used in this study was restricted to three for which The University of Waikato had previously obtained samples (Arawa-1, Ariki-1, and Kora-1). The fourth well, Wainui-1, obtained during the course of this study, was selected because of its more distal (basinward) position, sited on the edge of the modern day shelf break (Fig. 3.1). These four well sections, combined with data from the most recent biostratigraphic reports for wells in the study area, provide the basis for the chronostratigraphic framework constructed for the Giant Foresets Formation.

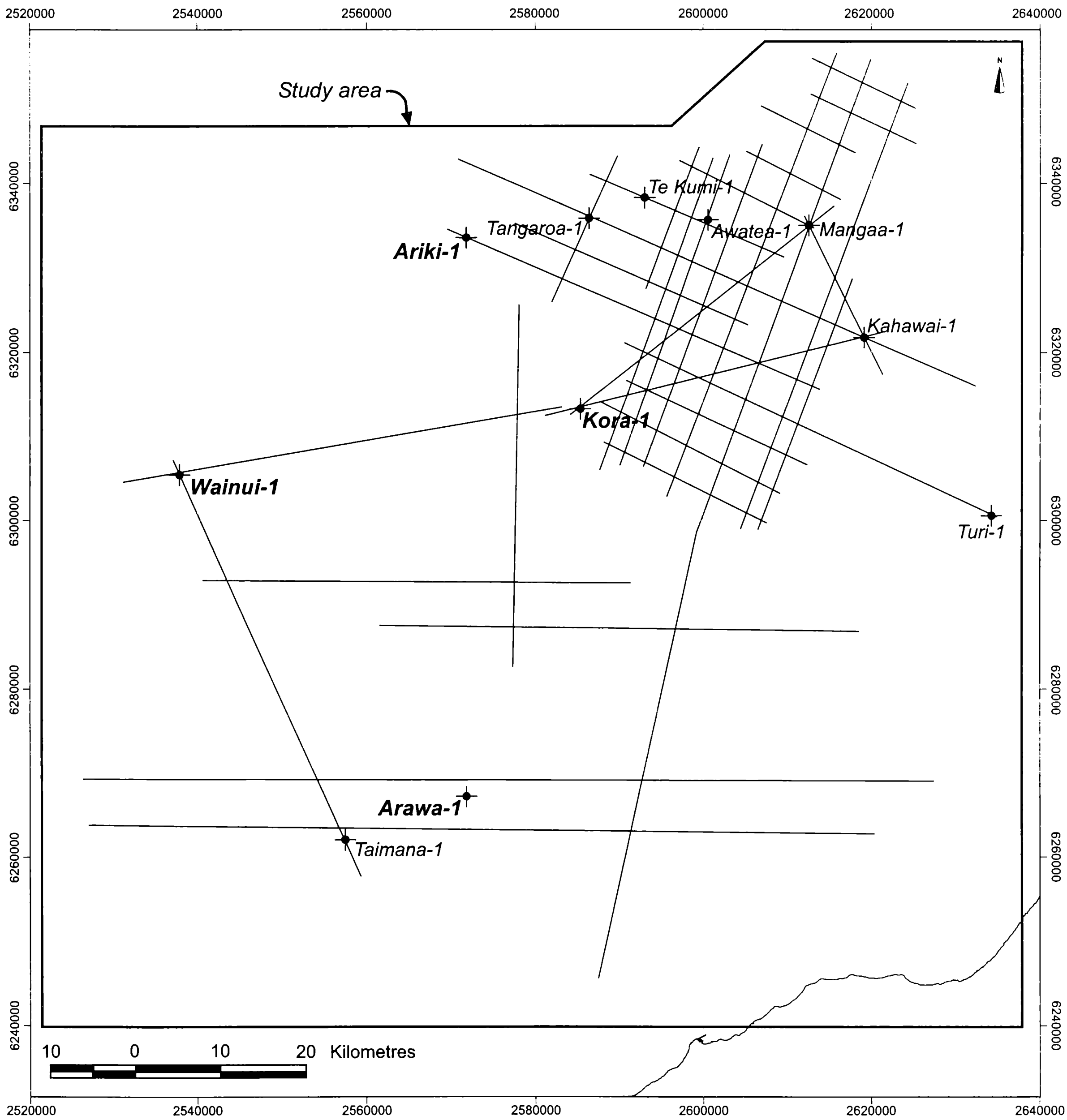


Fig. 3.1: Location map of the four well sections re-evaluated in this study (bold type) and other wells within the field area. Also shown are the seismic lines that have been used to correlate the biostratigraphy between wells.

3.2 Nomenclature and criteria used

A total of 378 cuttings samples from the four wells, spaced over approximately 5-15 m intervals, were picked for foraminifera. One hundred and seventy four of these underwent full benthic and planktic identification, while planktic versus benthic percentages were calculated for the remainder, and diagnostic globorotalids noted. The 150 μm to 2 mm size fraction was picked, eliminating juvenile specimens that are often difficult to identify. Species lists (proportional) are given in Appendix 3b, and sampled intervals for each well, together with total planktic specimens, are listed in Table 3.1. For each of the four wells discussed, cuttings are cited by their lower depth limit only. Following standards used in New Zealand Petroleum Reports, stage boundaries are located midway between the highest sample containing criteria for a new stage (first down-hole appearance or FDHA), and the lowest sample of the next overlying stage, unless otherwise stated. As with unpublished well reports, all care has been taken to identify cavings and contamination, which can be a major problem in some wells. Those foraminifera identified as not *in situ* have not been used in any biostratigraphic determinations. All depths quoted in this chapter are in driller depths (i.e., depth below Kelly Bushing or m bKB).

The lack of any absolute age control (i.e., isotopic or magnetic datums) means that a definitive chronostratigraphic framework for the Giant Foresets Formation cannot be constructed. However, although not an entirely satisfactory solution (see discussion in Nathan, 1997; Sherwood, 1997; Scott, 2001), bioevents are used as proxies for chronostratigraphic units. This study follows the standard New Zealand Cenozoic Stage classification of Morgans et al. (1997), with some modifications to the Mangapanian Stage, based on recent work undertaken by McIntyre (2001) and Kamp et al. (in prep.), and to the Waipipian Stage (after Kamp et al., in prep.). Chronostratigraphic ranges of key planktic fauna follow Scott et al. (1990) and Morgans et al. (1997) (Fig. 3.2).

There is some debate over whether New Zealand stages should be subdivided with the adjectives lower/upper, or early/late to denote position (e.g., Sherwood, 1997; Scott, 2001). While lower/upper is a preferred option by some (e.g., Scott, 2001), this study will continue with the use of early/late, partly because this usage is still dominant in literature, and partly because it is not within the scope of this study to discuss the issues related to chronostratigraphic and biostratigraphic classifications. Stage names within this chapter are frequently written in their abbreviated form, as listed below:

Wq	Haweran	0.33 – 0 Ma
Wc	Castlecliffian	1.6 - 0.33 Ma
Wn	Nukumaruan	2.28 - 1.6 Ma
Wm	Mangapanian	2.79 – 2.28 Ma
Wp	Waipipian	3.50 – 2.79 Ma
Wo	Opoitian	5.20 – 3.50Ma
Tk	Kapitean	6.60 – 5.20 Ma
Tt	Tongaporutuan	11.30 – 6.60 Ma

Stage divisions e – early, m – mid, l - late

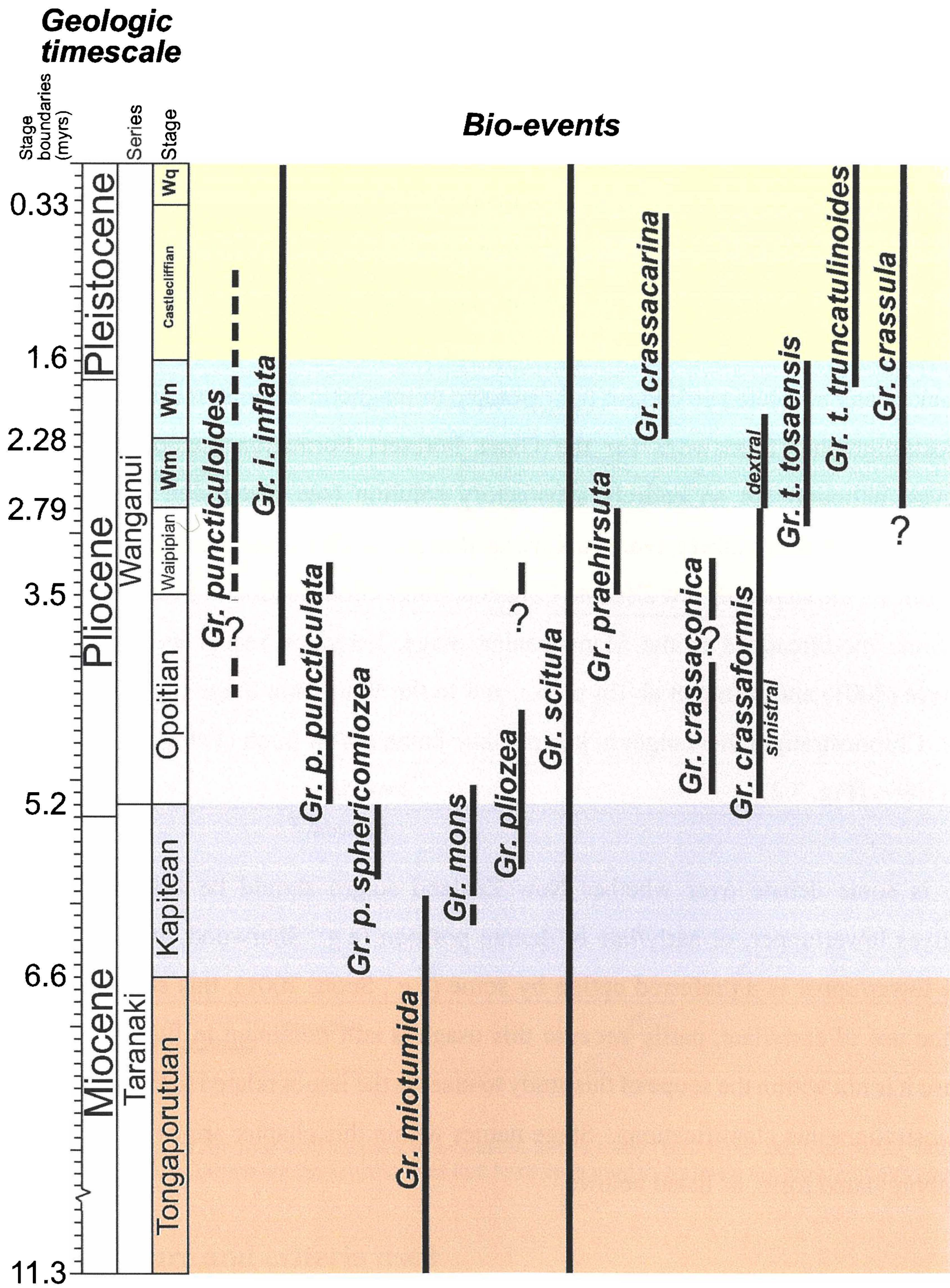


Fig. 3.2: Geological timescale and diagnostic (globorotalid) bioevents used in this study. Timescale is after Morgans et al. (1997) with revisions to the lower Nukumaruan, Mangapanian and Waipipian Stage boundaries after McIntyre (2001) and Kamp et al. (in prep.). Bioevents after Scott et al. (1990) and Morgans et al. (1997). Note that colours used for each stage are followed throughout this study. Wm, Mangapanian Stage; Wn, Nukumaruan Stage; Wq, Haweran Stage.

Table 3.1: Sample list, giving intervals picked for foraminiferal analysis, and total number of planktic specimens per interval.

Interval sampled	Arawa-1 Full ID	Arawa-1 Planktic %	No. planktic specimens
670-680 m	X		8
700-705 m	X		2
710-715 m	X		0
720-725 m	X		1
730-735 m	X		2
740-745 m	X		7
750-755 m	X		6
760-765 m	X		1
770-775 m	X		4
780-785 m	X		1
790-795 m	X		2
800-805 m	X		1
810-815 m	X		4
820-825 m		X	2
830-835 m		X	3
840-845 m		X	1
850-855 m	X		4
860-865 m		X	1
870-875 m		X	2
880-885 m		X	1
890-895 m...	X		2
1690-1695 m		X	135
1700-1705 m	X		79
1710-1715 m		X	118
1720-1725 m		X	157
1730-1735 m		X	218
1740-1745 m		X	170
1750-1755 m	X		128
	Total number of samples - 97		

Interval sampled	Ariki-1 Full ID	Ariki-1 Planktic %	No. planktic specimens
520 m	X		32
560 m	X		119
580 m		X	79
600 m	X		141
620 m	X		111
640 m		X	99
660 m	X		114
680 m		X	124
700 m	X		140
720 m		X	89
740 m	X		142
760 m		X	83
780 m	X		43
800 m		X	117
820 m	X		106
840 m		X	98
860 m	X		128
880 m		X	72
920 m	X		88
940 m		X	82
960 m	X		99
	Total number of samples - 89		

Interval sampled	Kora-1 Full ID	Kora-1 Planktic %	No. planktic specimens
630-640 m	X		13
640-650 m		X	9
660-670 m		X	39
670-680 m	X		1
680-690 m		X	19
700-710 m		X	19
710-720 m	X		6
720-730 m		X	9
740-750 m		X	43
750-760 m	X		14
760-770 m		X	21
780-790 m		X	4
790-800 m	X		2
800-810 m		X	12
820-830 m		X	11
830-840 m	X		7
840-850 m		X	11
860-870 m		X	9
870-890 m	X		2
880-890 m		X	9
900-910 m...		X	0
1670-1680 m	X		73
1680-1690 m		X	150
1690-1700 m	X		42
1700-1710 m		X	82
1710-1720 m	X		80
1720-1730 m		X	82
1730-1740 m	X		114
1740-1750 m		X	111
1760-1770 m		X	114
1770-1780 m	X		145
1780-1790 m		X	0
1790-1800 m	X		0
	Total number of samples - 102		

Interval sampled	Wainui-1 Full ID	Wainui-1 Planktic %	No. planktic specimens
470-480 m	X		0
480-490 m		X	137
500-510 m	X		67
520-530 m		X	100
540-550 m	X		88
560-570 m		X	55
580-590 m	X		49
600-610 m		X	39
620-630 m	X		102
640-650 m		X	85
660-670 m	X		43
680-690 m		X	29
700-710 m	X		44
720-730 m		X	51
740-750 m	X		52
760-770 m		X	71
780-790 m	X		39
800-810 m		X	31
820-830 m	X		49
840-850 m		X	40
860-870 m...	X		47
	Total number of samples - 90		

Full ID - samples that have undergone full benthic and planktic species identification
Planktic % - samples that have only had benthic:planktic ratios determined

Planktic foraminiferal species are most commonly used for age determinations in New Zealand, particularly the genus *Globorotalia*. Because of the frequent usage of this species in this chapter, the genus has been abbreviated to *Gr.* Benthic foraminiferal datums have been included in age determinations where identified.

3.3 Problems associated with well cuttings

Foraminiferal sampling from cuttings samples have a number of inherent limitations. Part of these limitations arise from the criteria imposed by the Ministry of Economic Development (Wellington) for well cuttings housed by Crown Minerals. Firstly, only sufficient sample to complete the analysis is allowed to be taken, and secondly, sampling is limited to no more than a third of the existing cuttings sample. For some wells with a diminishing stock of sample, only small samples for analysis can be obtained (Crown Minerals official policy, 2001). Consequently, after washing and sieving, a very muddy sample may yield only a small amount of usable material (in the 150 µm – 2 mm size fraction). In these samples, obtaining large foraminiferal populations is often impossible, even when the complete sample has been picked. All efforts have been made to advise the reader where specimen numbers are low, and identification of species tentative due to small population sizes. Relative abundances' (% planktics) on composite figures may be artificially high due to the low number of planktic specimens at some sites (e.g., if 2 of only 3 planktic specimens observed is *Gr. inflata*, imparting a relative abundance of 66%, a false impression of the dominance of this species is given). The reader should refer to Table 3.1 for actual species numbers.

Downhole contamination, through slumping and caving, is often a problem when dealing with cuttings samples, as is the homogeneity or degree of mixing of the sampled interval. Most cutting samples are composite in nature, and tend to incorporate the entire interval cited between sample depths. Unless a sample is labelled as a 'spot sample' (i.e., from a specified depth), it is assumed to be of a composite nature even if the sample bag is labelled with a single depth (M. Crundwell, 2001, pers. comm.). Samples are considered to be more reliable directly below cased intervals, and these have been marked on the composite figure for each well section analysed.

Many wells include intervals through which calcite encrustation is encountered. This occurs where sediment compaction and grain-to-grain pressure solution has distorted specimens so severely that confident identification is not possible. In several instances where calcite encrustation is prevalent, only approximate estimates of species numbers can be given.

3.4 Review of the biostratigraphy of Arawa-1, Ariki-1, Kora-1, Wainui-1, and Mangaa-1

The following section presents a diagnostic summary of each of the well sections reviewed by this study. Each stage is discussed in turn, and reasoning behind the allocation of stage boundaries given. Mangaa-1 is included in this review as it has been recently re-evaluated (King and Thrasher, 1996; Waghorn et al., 1996), and provides a fifth control well in the northeastern part of the study area (Fig. 3.1).

3.4.1 Arawa-1

Samples obtained from Arawa-1 are characterised by high percentages of calcareous benthic foraminifera relative to planktic species, and coarse, shelly sediment throughout much of the sampled interval, indicating a relatively shallow environment, and/or efficient transport from shallower water regions, resulting in dilution of planktic specimens (see Chapter 6 for detailed discussion). The low planktic numbers, coupled with a lack of diagnostic species, particularly in the upper 1400 m, have resulted in a limited ability to accurately constrain stages and stage boundaries. However, the increased number of samples examined in this study (97 between 670 and 1755 m) compared to the total examined by Crundwell et al. (1992; 18 over the same interval) has enabled subtle revision of the placement of the Opoitian and Nukumaruan to Castlecliffian Stage boundaries. Table 3.2 presents a summary of the previous biostratigraphy of Crundwell et al. (1992) and the revised biostratigraphy as determined by this study. The revised biostratigraphy is also illustrated in Fig. 3.3, which incorporates the relative abundances of each of the diagnostic species identified at this site. Sampled intervals occur in 5 m blocks, although generally only the lower depth limit is quoted (e.g., 945 m if the sample interval ranges between 940-945 m).

Table 3.2: Previous biostratigraphy of Arawa-1 (after Crundwell et al., 1992) and revised biostratigraphy from this study for the latest Miocene to Castlecliffian. Total depth of well is 3055 m.

Crundwell et al. (1992)		This study	
Depth (m)	**Stage**	**Depth (m)**	**Stage**
<668-980	e.Wn – Wc	670-717.5	Wc
980-1525	Wm – e.Wn	717.5-997.5	l.Wn
1525-1560	Wp	997.5-1525	Wm-e.Wn
1560-1760	Wo	1525-1567.5	Wp
1760-1960	l.Tk	1567.5-1715	Wo
1960-2486	l.Tt – e.Tk	1715-1760	e.Wo

Wanganui Series

Castlecliffian (670 – 717.5 m); Nukumaruan-Castlecliffian (717.5-887.5)

The upper boundary of the sampled interval cannot be determined due to lack of samples (first sample collected at 670-680 m). Planktic specimens occur sporadically and in low numbers (consistently less than 10% of total foraminifera, and fewer than 10 individuals), throughout this interval. *Globigerina* spp. dominates planktic species, with specimens of *Globorotalia inflata* (Wo-Recent) occurring sporadically. No other diagnostic planktics occur in this interval. Crundwell et al. (1992), identified *Gr. inflata triangula* (Wm-e.Wn) at 805 m, giving the interval 668 – 980 m an Early Nukumaruan to Castlecliffian age based on this and co-occurrences of the benthic species *Notorotalia zelandica* (Wm-Recent) and *Rotalia wanganuiensis* (Wn-Wc). *Globorotalia inflata triangula* is a species that Scott et al. (1990) consider to be a possibly short lived variant of *Gr. inflata*, and is not commonly recorded in New Zealand sequences. This study did not identify any 'typical' *Gr. inflata triangula* specimens, only confidently identifying *Gr. inflata* sensu stricto at this site. However, given the very low planktic counts, it is entirely conceivable that the former variant is present, but in very low numbers, and therefore not picked up by this study. Restriction to the Castlecliffian Stage is based on the identification of the benthic species *Cibicides marlboroughensis* (Recent) at 705 and 715 m, which confirm that the interval from 670 – 715 m is younger than Nukumaruan. However, given the lack of any useful Nukumaruan-restricted species, the rest of the interval is termed Nukumaruan to Castlecliffian based on the first downhole appearance (FDHA) of *Gr. puncticuloides* (l.Wp?/Wm-Wn; 1 specimen at 895 m, and sporadic occurrences thereafter).

Late Nukumaruan (887.5 – 997.5 m)

The Nukumaruan Stage is difficult to constrain at this site because of the paucity of diagnostic specimens. Benthic specimens cannot be used to help clarify the age of this interval, as most species are wide-ranging, spanning the Southland/Taranaki series to Recent. However, Crundwell et al. (1992) confirmed the Nukumaruan Stage with a tentative record of *Gr. crassula* (Wn-Rec) at 945 m. The entry of dextrally coiled *Gr. crassaformis* (Wm-e.Wn) is a well-defined event, confidently marking entry into Early Nukumaruan strata (Hornibrook, 1981). The FDHA of this species is observed at 1005 m, slightly higher than the previous authors, who identified *Gr. crassaformis* at 1015 m. Thus the interval 887.5 to 997.5 m is recognised as being younger than Early Nukumaruan, and therefore given the designation Late Nukumaruan. The event however is poorly resolved, again because of the rarity of the key species.

Mangapanian to early Nukumaruan (997.5-1525 m)

Planktic numbers steadily increase during this interval, but still comprise less than 25% of the total foraminiferal count. Low numbers and rarity of diagnostic species continue to make biostratigraphic-defined stages difficult to resolve. Diagnostic species in this interval include *Gr. puncticuloides*, morphotypes of *Gr. puncticulata* (Wo-Wp), and *Gr. crassaformis* (dextrally coiled), along with the ubiquitous *Gr. inflata*. The co-occurrence of *Gr. inflata* and *Gr. puncticuloides* from 895 – 1495 m, and persistent populations of dextrally coiled *Gr. crassaformis*, together with a lack of diagnostic Mangapanian-restricted species (the Mangapanian Stage is best defined on macrofaunal evidence), means that this interval can only be constrained to Mangapanian to Early Nukumaruan, consistent with Crundwell et al. (1992). The base of this interval is defined by the FDHA of *Gr. crassaconica* (Wo-Wp) at 1545 m, and interpolated between this sample and the next highest sample.

Waipipian (1525-1567.5 m)

A marked increase in planktic percentage occurs through this interval, with abundances rising to 90% of the total count at 1555 m, and a concomitant decrease in benthic specimens and diversity. *Globorotalia puncticulata* (Wo-e.Wp), common from 1515 m, begins to dominate globorotalid populations from 1545 m. This dominance coincides with the appearance of *Gr. crassaconica* (Wo-Wp) at 1545 m (identified in both this study and Crundwell et al. 1992), the benthic *Cibicides molestus* (Ab-Wp; Crundwell et al., 1992), and a change from dextrally coiled to dominantly sinistrally coiled populations of *Gr. crassaformis* (l.Tk-Wp). Together, these events reliably indicate entry into Waipipian strata. *Gr. inflata* is still common throughout this interval (to 1565 m), and persists in sporadic and low numbers to 1615 m.

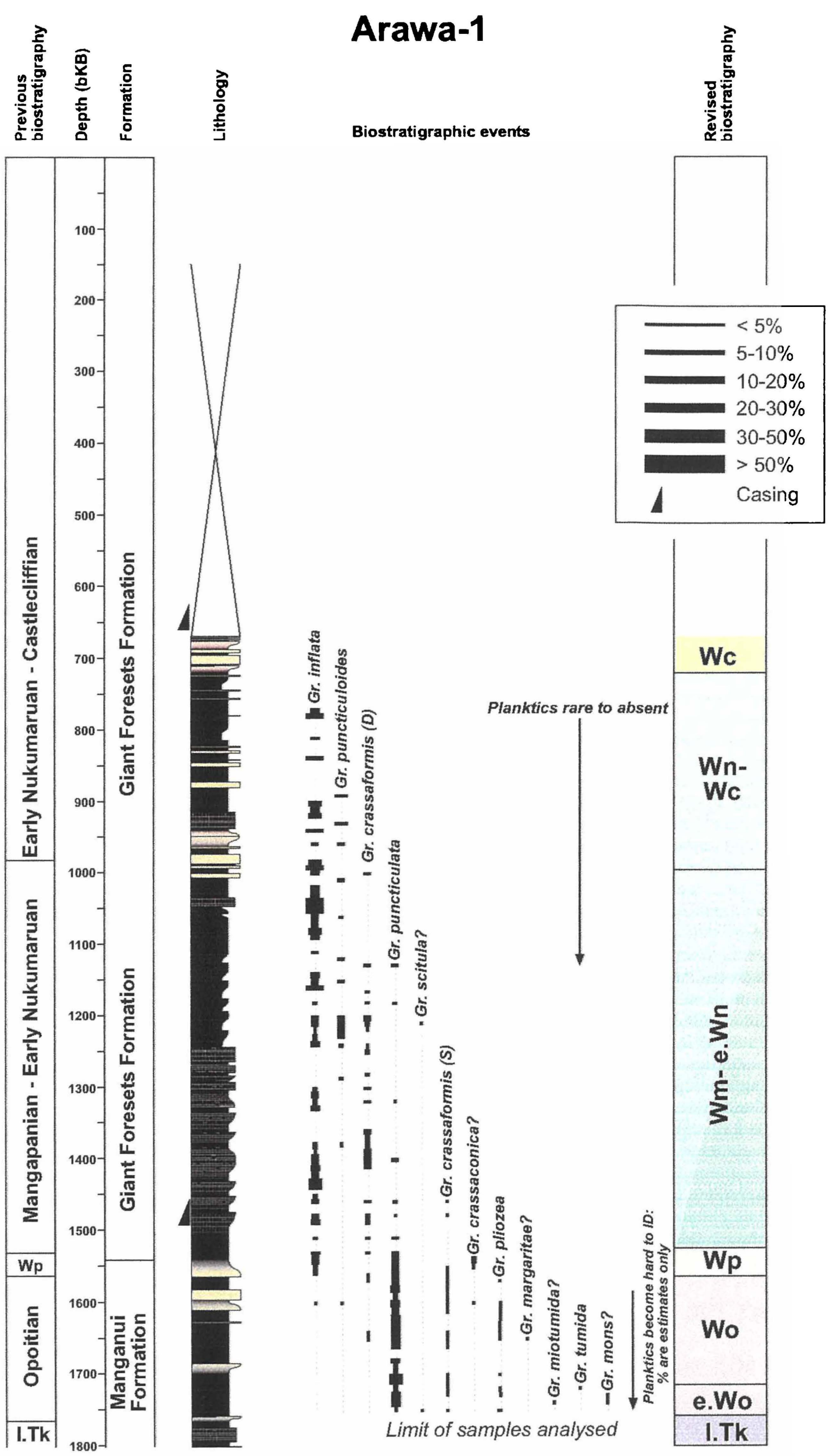


Fig. 3.3: Revised biostratigraphy for Arawa-1, to 1750 m. Previous biostratigraphy after Crundwell et al. (1992).

The base of the Waipipian Stage is poorly constrained at this site. However, the presence of *Gr. pliozea* (l.Tk-l.Wo) at 1575 m suggests entry into the Opoitian Stage. Coupled with a change in populations dominated by *Gr. inflata* to populations dominated by *Gr. puncticulata* at approximately the same sample depth, the lower boundary for this stage is interpolated between the FDHA of *Gr. pliozea* at 1575 m, and the previous sample at 1565 m.

Opoitian (1567.5-1760 m)

Foraminifera are abundant and are initially well preserved in the upper part of this interval. However, below 1605 m poor preservation of foraminifera make species increasingly difficult to identify. *Globorotalia inflata* persists until approximately 1615 m, at which depth it abruptly disappears from the foraminiferal record. *Globorotalia puncticulata* becomes the dominant globorotalid instead. This suggests the sequence below 1565 is at least Opoitian in age. *Globorotalia pliozea* also persists to the base of the interval, occurring together with *Gr. mons* (Tk-e.Wo; first noted by Crundwell et al., 1992, at 1716 m, and by this study at 1735 m), and strongly sinistral populations of *Gr. crassaformis*, confirming an Opoitian age. The presence of *Gr. mons* at 1716 m (Crundwell et al., 1992), and 1735 m (this study), but not at 1715 m, indicates that the interval between ~ 1715 m and 1760 m is Early Opoitian. Rare specimens of *Gr. miotumida* (l.Sl-m.Tk) found at 1745 m, while in good condition, are considered to be reworked, as all other specimens confirm an age no older than Opoitian.

The upper boundary of the Opoitian Stage in this study is delineated by the FDHA of *Gr. pliozea*. This differs from Crundwell et al. (1992) who used a shift in population in the *Gr. inflata* lineage from specimens with 4 chambers in the outer whorl (*Gr. puncticulata*) to specimens with 3-3.5 chambers (*Gr. inflata* s.s) to locate top Opoitian strata. In the New Zealand region this gradational transformation occurs in the Late Opoitian (Scott et al. 1990) with *Gr. inflata* dominant from this stage to Recent. Using this criterion, Crundwell et al. (1992) indicate that this shift occurred between 1542-1578 m (interpolated to 1560 m). This study instead prefers the previous option because of the relatively low numbers of planktic specimens above 1565 m and therefore the limited ability to observe this 'population shift'. Both these methods, however, have resulted in approximately the same location for the upper Opoitian boundary.

Taranaki Series

The lowest sample picked in this study is 1755 m, therefore biostratigraphic stages below this depth follow those of Crundwell et al. (1992).

Unconformities

No unconformities are indicated in either the well completion report (ARCO Petroleum, 1992), or the paleontological report (Crundwell et al., 1992). It should be noted, however, that the base of the Giant Foresets Formation (at 1541 m) approximately correlates with a dramatic increase (downhole) in planktic percentage noted between 1530 and 1540 m. This downhole increase is also noted at a number of other well sites within the study area (discussed fully in Chapter 6), where it is often associated with a change downwards in sediment type from fine-grained lithologies of the Giant Foreset Formation, to the marly and highly calcareous lithology that characterises the Ariki Formation. On wireline logs, there is often an observable 'kick' at this stratigraphic level (refer to Chapter 4) and this frequently corresponds with a seismically distinctive (bold) horizon (refer to Chapter 5).

3.4.2 Ariki-1

At the time of drilling, no diagnostic foraminiferal work was undertaken for this site. Instead, Shell BP Todd Oil Services (1984) applied their biostratigraphy from neighbouring Tangaroa-1 to Ariki-1. A review of the biostratigraphy of Ariki-1 was later undertaken as part of the Cretaceous-Cenozoic Project (CCP) in west Taranaki (Hayward, 1986b), with ages further refined by King and Thrasher (1996) in light of new understanding regarding planktic lineage's and associated stages. Previous studies evaluated only 13 samples from the upper (Plio-Pleistocene) part of the sequence drilled at this site. In contrast, this study has analysed a total of 77 samples from the Giant Foresets Formation, a marked increase in sampling resolution, with a further 12 samples through the Ariki Formation. This has resulted in some dramatic shifts to stage boundary positions, as summarised in Table 3.3 and Fig. 3.4.

Wanganui Series

Late Nukumaruan – Castlecliffian (490-900 m); Late Nukumaruan (900-1390 m)

Foraminiferal assemblages in samples are diverse, becoming more so with depth. Planktics are well preserved, although many benthic species throughout this interval (e.g., *Quinqueloculina*) display evidence of transport (i.e., abrasion, fragmentation). Planktic percentages vary from 10 to 85%, averaging 50-60%, reflecting the more distal (basinward) position of this site. Benthic specimens tend to be large and robust, and planktic specimens, particularly *Gr. inflata*, very inflated and globose, indicating more oceanic conditions than experienced at Arawa-1.

Table 3.3: Previous biostratigraphy of Ariki-1 (after Hayward, 1986b; King and Thrasher, 1996) and revised biostratigraphy from this study for the latest Miocene to Castlecliffian. Total depth of well is 4822 m.

Hayward (1986b) and King and Thrasher (1996)		This study	
Depth (m)	**Stage**	**Depth (m)**	**Stage**
474 - 850	l.Wn-Wc	490-900	l.Wn-Wc
850-1375	l.Wn	900-1390	l.Wn
1375-1900	e.Wn	1390-1810	e.Wn
1990-2000	Wp	1810-2100	Wp
2130-2140	l.Wo	2100-c.2153	l.Wo
c.2153	m.Wo	c.2153	m.Wo
2170-2175	m.Tk		
2240-2320	l.Tt-m.Tk		
2350-2355	e.Tt		

Globorotalia inflata (l.Wo-Recent), as with most well sections, is ubiquitous throughout most of the sampled interval. This species is common in facies deeper than about middle shelf, and is the most widely recorded globorotalid from the upper Wanganui Series (Scott et al., 1990). At this site, *Gr. inflata* occurs in good numbers, together with *Gr. truncatulinoides* (l.Wn-Wc) above 1380 m, indicating that the stratigraphic interval above 1380 m is no older than Late Nukumaruan. Sporadic occurrences of *Gr. truncatulinoides* below 1380 m (e.g., at 1680 and 1760 m) are noted, though these appear to be intrapopulational variants between *Gr. t. tosaensis* (Wn-Wn) and *Gr. truncatulinoides* (Hornibrook, 1981). Scott et al. (1990) state that morphotypes of the latter may closely resemble the former, with a polygonal spiral outline and a weak peripheral keel.

The presence of *Cibicides neoperforatus* (Tt-Wn) below 920 m, *Bolivina* cf. *affiliata* (l.Pl-Wn), and *Cibicides deliquatus* (l. Sl-e.Wn), at 1000m, *Uvigerina pliozea* (Tt? Wo-Wn) below 1000 m, *Karreriella cylindrica* (Pl-Wn) at 1040m, together with *Gr. truncatulinoides* to 1380 m, gives a Late Nukumaruan to Castlecliffian age for the interval 490 – 900 m, and a Late Nukumaruan age for the interval 900 –1390 m. This follows the scheme of Hayward (1986b), but raises the boundary for entirely Nukumaruan strata to 900 m, compared with 1000 m in the earlier report.

Early Nukumaruan (1390-1810 m)

Gr. inflata continues to dominate throughout this interval, comprising up to 37% of planktics, while *Gr. truncatulinoides* is only rarely observed. The general absence of this latter species below 1380 m, and the presence of strong populations of *Gr. crassula* (Wn-Wc) above 1820 m, and *Gr. crassacarina* (Wn-Wc) above 1760 m (1800 m in Hayward, 1986b), gives an Early Nukumaruan age for the interval 1380 – 1810 m.

Dextral and sinistral specimens of *Gr. crassaformis* occur sporadically throughout the sequence, and neither coiling direction dominates. Because of the very intermittent occurrence and low numbers of dextral populations of *Gr. crassaformis*, which is often used to delineate the Mangapanian Stage, it is almost impossible to identify Mangapanian strata in this sequence. *Globorotalia puncticuloides* (upper Wp? – Wm/Wn) is a species that could be used to help define the Mangapanian Stage, but at this site, occurrences do not extend below the *Gr. crassula* zone. It is unlikely that the Mangapanian Stage is absent from the sedimentary sequence at this site, only unable to be recognised because of a distinct lack of diagnostic criteria. There may be some precedent for including Mangapanian strata between the *Gr. crassacarina/Gr. crassula* zone and the first occurrence of *Gr. puncticulata* (Wo-e.Wp) at 1900 m, though for now the Mangapanian Stage is not recognised.

Waipipian (1810-2100 m)

Wide spacing of samples in the study undertaken by Hayward (1986b) resulted in a decreased ability to constrain stages and stage boundaries. This study has been able to refine upper and lower stage boundaries because of the higher sample resolution, though some problems have arisen where foraminifera were found to be stratigraphically out of place, apparently not due to contamination. For example, *Gr.* aff. *praehirsuta* is a globorotalid restricted to the Waipipian, and therefore is considered an excellent discriminator of this stage. In Ariki-1, this species is found higher in the sequence than one would expect. The first occurrence, noted at 1420 m, is only very tentative, and therefore not much confidence is placed on it. Other occurrences involve only rare specimens, but identifications are more confident. It may be that this species actually does extend higher in the biostratigraphic record than previously recorded, or these specimens *have* actually been reworked from older sequences and transported to this site. However, while good populations of *Gr.* aff. *praehirsuta* do not occur at this site, they are most numerous below 1810 m.

The Waipipian Stage at this site is based on several diagnostic criteria other than *Gr.* aff. *praehirsuta*. These include the absence of *Gr. crassula*, *Gr. crassacarina*, and *Gr. puncticuloides* below 1820 m, confidently indicating entry into Waipipian strata, and the (abundant) presence of *Gr. inflata* without the presence of *Gr. puncticulata* (Wo-e.Wp) throughout much of this interval. Hayward (1986b) also notes the presence of *Gr. subconomiozea* (Wp) at 1900 m, and the absence of the benthic species *N. zealandica* (Wm-Rec) below 1700m. The FDHA of *Gr. puncticulata* at 1980 m marks entry into the basal (early) Waipipian Stage. The base of this interval, at 2100 m, is defined by the first occurrence of *Gr. pliozea* (l.Tk-l.Wo) at 2110, and is interpolated between this interval and the preceding sample.

Opoitian (2100-ca. 2153 m)

Persistent populations of *Gr. inflata* down to 2140 m, with the co-occurrence of *Gr. pliozea* and the absence of other criteria, suggest that the interval is entirely lower Opoitian in age to at least this depth. The presence of the benthic species *Bolivinita* aff. *pohana* (Wo) at 2100 m further confirms entry into Opoitian strata. While the first appearance datum (FAD) of *Bolivinita pohana* in sedimentary sequences is most often used to delineate the Tongaporutuan-Kapitean boundary, the form called *B.* aff. *pohana* is found in the Kapitean to Opoitian (Hornibrook et al., 1989). Both this study and that of Hayward (1986b) note the sudden absence of *Gr. inflata* at 2150 m. Large populations of *Gr. puncticulata* without *Gr. inflata* suggest that the very lowest part of this interval is mid Opoitian, in agreement with Hayward (1986b).

Hayward (1986b) notes that the sample at 2150 m contains elements of deeper water benthic fauna, indicating a distinct change from the shallower water assemblages found previously. This sample is inferred to cross an unconformity between Wanganui Series strata and Taranaki Series strata (see below). While it does not contain any diagnostic Taranaki Series fauna, there is a radical change in benthic fauna. Because of this, Hayward (1986b) placed part of this sample in the Wanganui Series, and part in the lower Taranaki Series. This scheme is retained here.

Taranaki Series (> 2153 m)

A distinct change in population, as well as the benthic versus planktic ratio, is observed between the sampled interval at 2140 m (less than 50% planktics) and the sampled interval at 2150 m (greater than 80% planktics). The elevated planktic to benthic ratio corresponds to the radical faunal change noted earlier, with deeper water assemblages continuing into the lower Taranaki and Pareora Series (Hayward, 1986b). Faunas are moderately well preserved, but display pressure solution and compaction due to sediment loading, making confident identification of species difficult.

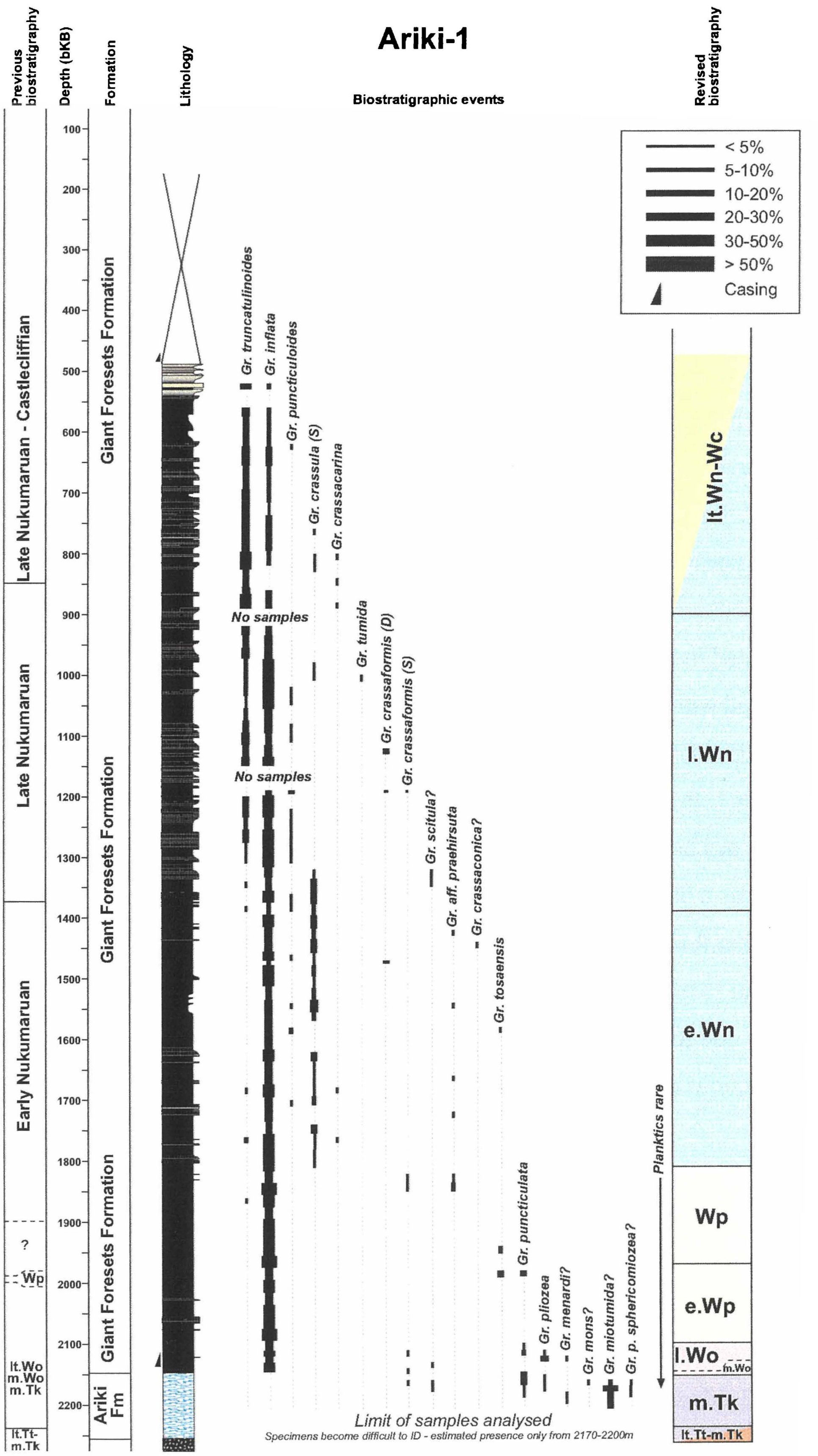


Fig. 3.4: Revised biostratigraphy for Ariki-1, to 2150 m. Previous biostratigraphy after Hayward (1986b), and King and Thrasher (1996).

The short-lived *Gr. puncticulata* is abruptly replaced at 2160 m by numerous other species including *Gr. miotumida* (m.Tk; becomes dominant), *Gr. p. sphericomiozea* (l.Tk), and *Gr. pliozea* (possibly caved). Specimens of *Gr. puncticulata* are probably caved. The presence of *Gr. p. sphericomiozea* and *Gr. conomiozea* (Tk), noted by Hayward (1986b) at 2170 m, together with *Gr. miotumida*, give a mid to (Late?) Kapitean age for the interval to 2200 m. Hayward (1986b) gives a Late Tongaporutuan to mid Kapitean age for the interval 2240-2300, and an Early Tongaporutuan age for 2350 m.

Unconformities

The distinct change in faunal assemblage and planktic-benthic proportion between 2140 and 2150 m, observed in both this study and that of Hayward (1986b), suggests the presence of an unconformity. Although the sample at 2150 m contains no distinctly Taranaki Series specimens, it does contain elements of distinctly deeper water benthic faunas compared to the assemblage observed at 2140 m (Hayward, 1986b). Given that the sample at 2140 m is definitely placed in the Wanganui Series, and the sample at 2160 m in the Taranaki Series (m-?l. Tk), coupled with evidence of faunal change at 2150 m, a time gap (or possibly an extremely condensed interval) within this latter sample is indicated, with ?Late Kapitean to Early Opoitian strata missing. In the well completion report (Shell BP Todd Oil Services, 1984), this unconformity has been located at 2153 m, near the top of the Ariki Formation, and is picked on wireline logs by a discernable shift in all log traces (Chapter 4). While Mangapanian strata are not identified at this site, it is highly probable that it *is* represented within the Giant Foresets Formation, but is unable to be recognised because of an absence of diagnostic species at this site, possibly due to environmental conditions. Wireline logs and lithological characteristics suggest continuous sedimentation, with no distinct breaks noted.

3.4.3 Kora-1

Kora-1 was drilled into the side of the Kora volcanic structure, targeting possible reservoirs on the flanks of the submarine volcano (ARCO Petroleum NZ Inc., 1988). Foraminiferal analyses were initially undertaken by Hayward and Strong (1988), who examined a total of 11 cuttings samples and 2 sidewall cores in the post volcanic part of the sequence. Two fill-in samples were later examined by Waghorn et al. (1996) to try to refine biostratigraphy of the Tongaporutuan to Nukumaruan Stage tops. This study analysed a total of 102 samples through the Giant Foresets Formation and very upper part of the Miocene volcanic complex. As summarised in

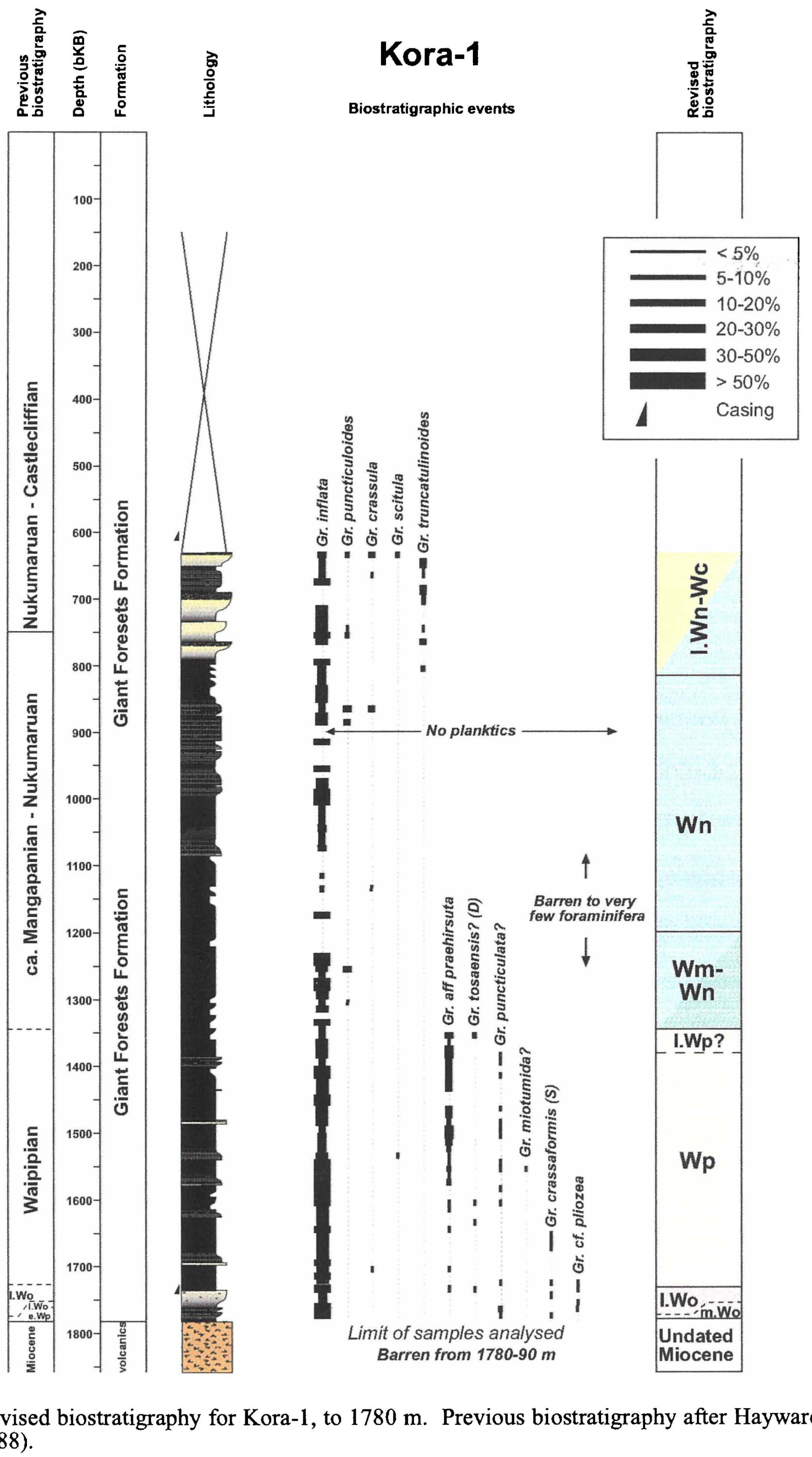


Fig. 3.5: Revised biostratigraphy for Kora-1, to 1780 m. Previous biostratigraphy after Hayward and Strong (1988).

Nukumaruan (815-1200); Mangapanian-early Nukumaruan (1200-1345)

Globorotalia inflata persists throughout this interval, together with the occasional *Gr. puncticuloides* (above 1310 m), rare *Gr. crassula* (above 1160 m), and moderate to large

Table 3.4 and Fig. 3.5, higher resolution sampling resulted in considerable changes to the previous biostratigraphy, particularly redefinition of the Nukumaruan and Mangapanian Stage boundaries, and subtle redefinition of the position of the (lower) Waipipian Stage boundary.

Table 3.4: Previous biostratigraphy of Kora-1 for the Giant Foresets Formation (after Hayward and Strong, 1988) and revised biostratigraphy from this study. Total depth of well is 3421 m.

Hayward and Strong (1988)		This study	
Depth (m)	**Stage**	**Depth (m)**	**Stage**
630-700 m	Wn-Wc	630-815	1.Wn-Wc
800-1300 m	ca. Wm-Wn	815-1200	Wn
1390-1700 m	Wp	1200-1345	Wm-e.Wn
1767 m	l.Wo	1345-1730	Wp
1779.6 m	m-l.Wo boundary	1730-1770	l.Wo
1960-2486	l.Tt – e.Tk	1779.6	m.-l.Wo

Wanganui Series

While foraminiferal faunas are well preserved throughout the sampled interval, they are highly variable in abundance. Planktic numbers are especially low, often less than 20 specimens in a sample, and less than 20% of the total faunal count. From approximately 1390 m downwards, planktic numbers increase, comprising up to 60% of the total count. Samples between 1090 and 1260 m are either barren of, or have very few foraminifera, particularly age-diagnostic species.

Late Nukumaruan - Castlecliffian (630-815 m)

Because of low specimen numbers and lack of age-diagnostic planktic or benthic species at this site, more detailed sampling has done little to refine the previous biostratigraphic scheme. *Globorotalia inflata* is the dominant planktic throughout the Giant Foresets Formation, often comprising greater than 30% of planktics. It co-occurs with sporadic observations of *Gr. truncatulinoides* (l.Wn-Recent), *Gr. crassula* (Wn-Recent) and *Gr. puncticuloides* (l.Wp? Wn-Wc), indicating Late Nukumaruan to Castlecliffian strata. The base of this interval has been interpolated between the lowest occurrence of *Gr. truncatulinoides*, at 810 m, and the next lowest sample.

populations of the benthic foraminifer *Notorotalia zelandica* (Wm-Recent) above 1050 m. As with so many well sections in Taranaki basin, it is very difficult to pick the Mangapanian Stage because of a lack of distinctive fauna, and at this site in particular because of the general paucity of fauna obtained from cuttings samples. The lowest occurrence of *Gr. crassula* at 1140m suggests that the interval above this could be restricted to the Nukumaruan Stage. Lack of fauna allows this boundary to be placed lower, and a position half way between 1140 m and 1260 m is arbitrarily chosen as the base of entirely Nukumaruan strata.

Occurrences of *Cibicides deliquatus* (l.Pl-e.Wn) below 1290 m imply that strata at this depth (downhole) are no younger than Early Nukumaruan, while the FDHA of *Gr.* aff. *praehirsuta* (Wp) at 1360 m confidently heralds entry into the Waipipian Stage. Through lack of any other criteria, the interval between 1185 and 1345 m is interpreted as Mangapanian to Early Nukumaruan in age.

Waipipian (1350-1730 m)

While no diagnostic planktic foraminifera occur between 1310 and 1350 m, a sudden influx of *Gr.* aff *praehirsuta* (Wp) at 1360 m (comprising up to 20% of planktics) occurs. This places the upper boundary for this stage at approximately 1345 m. Regular occurrences of *Gr. puncticulata* (Wo-e.Wp), rare observations of *Gr. t. tosaensis* (l.Wp-Wn) and sinistrally coiled specimens of *Gr. crassaformis* (l.Tk-Wp; Waghorn et al., 1996) confirm a Waipipian age. A rare and heavily encrusted specimen of *Gr. miotumida* (1550m; tentative identification) appears to be reworked and is thus discounted from age determinations. The base of this stage is interpolated to be immediately below the last downhole appearance (LDHA) of this species at 1740 m.

Late Opoitian (1730-1780 m)

The interval from 1750 – 1790 m is almost entirely Late Opoitian in age, based on the absence of *Gr.* aff. *praehirsuta* and the presence of *Gr. inflata*. Occasional specimens of *Gr. puncticulata* occur between 1770 and 1790 m, although these appear to be intermediate forms, such as encountered by Hayward and Strong (1988) in a sidewall core at 1779.6 m.

The presence of *Gr. subconomiozea* (m.Wo; after Hayward and Strong, 1988) at 1779.6 m, and the co-occurrence of *Gr. inflata*, gives a Late Opoitian age for this sample. No foraminifera were found in cuttings samples at 1790 and 1800 m, both compositionally distinct (volcaniclastic) from overlying cuttings samples, and barren of foraminifera, suggesting that the basal part of the Opoitian is absent from the stratigraphic record. The Kora volcanic complex has been dated only as undifferentiated Miocene in age (to 2580 m). The biostratigraphic scheme below 1790 m follows that of Hayward and Strong (1988).

Unconformities

An unconformity appears to exist between the top of the Miocene volcanic complex, and the overlying mid-Late Opoitian strata. Numerical dating at the base of the volcanic complex reveals an age of approximately 17 Ma (Mid Altonian), and suggests that Early Miocene to Early Pliocene strata are absent (Hayward and Strong, 1988).

3.4.4 Wainui-1

Wainui-1 is situated at the edge of the modern day shelf break and is the second-most westerly well drilled to date in the Taranaki Basin (with only Tane-1 being situated further to the west). The original well completion report (Shell BP Todd, 1982) did not have available to it the results of then recently published work by Hornibrook (1981, 1982) or the unpublished work of G. Scott (IGNS). With this in mind, Hayward (1984) re-examined the original slides, including 18 cuttings samples and one sidewall core from the Giant Foresets and Ariki Formations. Slides were reassessed by Wonders (1986) as part of an integrated report assessing Wainui-1 and Tane-1. As with Ariki-1, stages presented in King and Thrasher (1996) follow those of Hayward (1984), with recent adjustments. This study has examined a total of 90 samples, the majority of these from the Giant Foresets Formation, and two from the very top part of the Ariki Formation. The composite results of the previous studies, and the biostratigraphy as determined by this study are presented in Table 3.5 and Fig. 3.6.

Wanganui Series

Planktic numbers vary considerably through Wanganui Series strata, from less than 5% to over 90%, and actual specimen numbers are relatively high, reflecting the more basinward position of this well site. A number of barren zones exist, including the first cuttings sample at 480m (coarse sandstone) and an interval between 1615 and 1655 m. *Globorotalia inflata* dominates planktic species, comprising 15 to 70% of total planktics.

Table 3.5: Previous biostratigraphy of Wainui-1 (after Hayward, 1984; Wonders, 1986; King and Thrasher, 1996) and revised biostratigraphy from this study, latest Miocene to Nukumaruan. Total depth of well is 3879 m.

Hayward (1984), Wonders (1986) and King and Thrasher (1996)		This study	
Depth (m)	**Stage**	**Depth (m)**	**Stage**
500-710	l.Wn	480-795	l.Wn
800-1450	Wn	795-1535	Wn
1450-1550	Wm	1535-1645	Wm
1550-1950	Wp	1645-1840	Wp
1950-2210	l.Wo-Wp	1840-2110	Wo - Wp
2218	l.Tt-m.Tk	2110-2210	l.Wo
2300-2410	l.Tt	2210-2250	m.Tk
2500-2510	e.Tt		

Late Nukumaruan (480-795 m)

Common occurrences of *Gr. truncatulinoides* (l.Wn-Recent) above 790 m, together with *Gr. crassacarina* (Wn-Wc), *Gr crassula* (Wn-Recent), and the ever-present *Gr. inflata* (l.Wo-Recent) initially give a Late Nukumaruan to Castlecliffian age for the interval 470–740 m bKB. However, use of some age-restricted benthic species, such as *Cibicides neoperforatus* (Tt-Wn, above 910m), *Uvigerina pliozea* (Wo-Wn), *Rotalia wanganuiensis* (Wn-Wc, above 770 m), and *Plectofrondicularia pellucida* (Tt-Wn; noted at 600 m by Hayward, 1984), enable this interval to be constrained to Late Nukumaruan. While Wonders (1986) inferred a Late Nukumaruan-Recent age for this interval based on the presence of *Notorotalia finlayi* (Wp-Recent), *N. zelandica* (Wm-Recent) and *Bolivinita quadrilatera* (Tk-Recent), this study prefers the former option, which also concurs with Hayward (1986), recognising that while these aforementioned species range into the Recent, the presence of the Nukumaruan-restricted benthic and planktic species suggests that this interval is at youngest, Late Nukumaruan. The base of this interval is interpolated between the LDHA of *Gr. truncatulinoides* (790 m) and the next lowest sample.

Nukumaruan (795-1535 m)

Globorotalia crassula occurs persistently above 1530 m, together with *Gr. crassacarina* (above 1530 m). Assemblages are consistently dominated by *Gr. inflata*, though sporadic occurrences of *Gr. puncticuloides* (l.Wp? Wm-Wn) are also noted throughout this interval. Notable benthic species include persistent populations of *Uvigerina pliozea* and occasional *Cibicides neoperforatus* (Tt-Wn), confirming a Nukumaruan age.

The base of the Nukumaruan Stage as determined in this study is lower than indicated in both Hayward (1984; gives a Wm age to a sample at 1500 m) and Wonders (1986; suggest a Wn-Wm age for 1000-1410 m, and a Wm age for 1500-1510 m). Hayward (1984) based a Mangapanian age on the disappearance of *Gr. crassula*, and the appearance of the benthic species *Notorotalia taranakia* (range given as Sw-Wm in that report). *Globorotalia crassula* was identified at 1530 m by this study, appearing to be *in situ* rather than reworked; this study also assigns the age range for *N. taranakia* given in Hornibrook et al. (1989) of Waiauan to Nukumaruan. The FDHA of *N. taranakia* is observed at 550 m *(N.* cf. *taranakia*), and is consistent with a Nukumaruan age or older.

Mangapanian (1535-1645 m)

The Mangapanian Stage is difficult to identify in the Plio-Pleistocene sequence at this site. Hayward (1984) and Wonders (1986) both delineated the Mangapanian Stage on the presence of *Gr. tosaensis* (l.Wp-e.Wc), the disappearance of *Gr. crassula*, and the first *in situ* occurrence of *Notorotalia taranakia.* As mentioned above, this study has found *in situ Gr. crassula* lower down in the sequence, and uses a different age range for *N. taranakia.* The presence of *Gr. t. tosaensis* between 1550 and 1850 m, the disappearance of *Gr. crassula* between 1530 and 1550 m, the last apparently *in situ* occurrence of *N. zelandica* (Wm-Recent) at 1710 m, and the appearance of good populations of *Gr.* aff. *praehirsuta* (Wp) at 1690 m, but lack of faunas from 1615 to 1655 m, puts the lower boundary of the Mangapanian Stage between 1690 m and somewhere within the barren zone. Interpolation places the boundary at approximately 1645 m.

Waipipian (1645 - 1840); Opoitian to Waipipian (1840- 2110 m)

Globorotalia aff. *praehirsuta* (Wp) is relatively persistent from 1690 m to 2110 m. Although planktic numbers are low from 1810–2210 m (refer Table 3.1) and consequently species percentages are inflated, the presence of this species, and co-occurrence of *Gr. t. tosaensis* (above 1850 m) and *Gr. puncticulata* (Wo-e.Wp; FDHA at 1760 m) indicate a Late Waipipian age for the interval between 1645 and 1850 m, and an Opoitian to Waipipian age between 1850 and 2110 m. This is generally consistent with Hayward (1984), who gives the interval 1600-2210 m a Late Opoitian to Waipipian age.

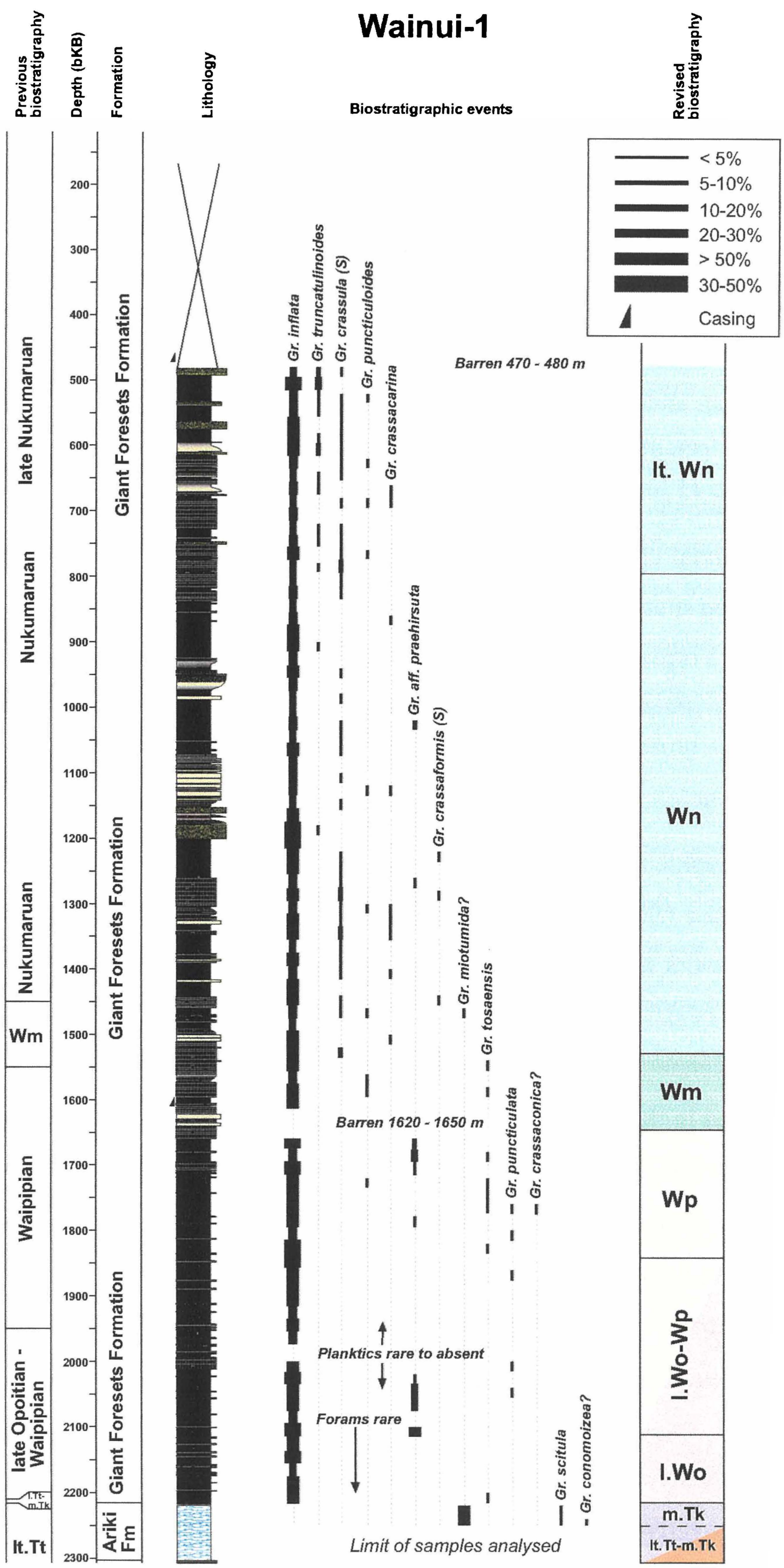


Fig. 3.6: Revised biostratigraphy for Wainui-1, to 2250 m. Previous biostratigraphy after Hayward (1984), Wonders (1986), and King and Thrasher (1996).

Late Opoitian (2110-2210 m)

A tentative identification of a specimen of *Gr.* aff. *praehirsuta* at 2110 m, the absence of this species at 2130 m, and the continued presence of *Gr. inflata* to 2210 m suggest that the interval between 2110 and 2210 can be no older than Late Opoitian. Wonders (1986) suggested that there is no evidence to indicate the presence of any Wo strata.

Taranaki Series

Globorotalia inflata (l.Wo-Recent) is abruptly replaced by *Gr. miotumida* (l.Sl-m.Tk) at 2230 m, suggesting that an unconformity exists between 2210 and 2230 m. Analysis of a sidewall core from 2218 m (Shell BP Todd, 1982) reveals a Late Tongaporutuan to mid Kapitean foraminiferal assemblage, confirming the presence of an unconformity between 2210 and 2218 m. This sudden faunal change is mirrored by a correspondingly abrupt change in lithology, from the fine-grained siltstone and sandstone of the Giant Foresets Formation, to the calcareous marl of the Ariki Formation, and a corresponding shift in wireline signatures. Planktic percentages increase dramatically also, comprising greater than 80% of the total faunal count at 2230 m.

Previous biostratigraphic reports (Shell BP Todd, 1982; Hayward, 1984) identified the planktic species, *Gr. conoidea* (Sl-e.Tt) as the dominant globorotalid species below 2210 m. More recently published work has synonymised *Gr. conoidea* with *Gr. miotumida* (e.g., Scott et al., 1990). Hornibrook et al. (1989) refers thicker-walled, 5-chambered forms to *Gr. conoidea*, and thin-walled (after Kennett and Srinivasan, 1983) 4-chambered variants to *Gr. miotumida*. All specimens identified in this study are 4-chambered, and show little calcite thickening, and are therefore referred to *Gr. miotumida*. This species is the dominant globorotalid species between 2210 m and 2250 m (stratigraphically lowest sample analysed by this study), and together with *Gr. conomiozea* (Tk) confirms a mid Kapitean age to approximately 2250 m. Biostratigraphic zones below 2250 m follow those applied by Hayward (1984).

Unconformities

As suggested previously, a time gap is indicated between 2210 and 2230 m, between the Ariki Formation and the overlying Giant Foresets Formation. Hayward (1984) suggested that all of the Opoitian, and most (or all) of the Kapitean is missing. This differed from Wonders (1986), who, using a different stratigraphic range for the benthic species *Uvigerina notohispida*, identified the sidewall core at 2218 m as being only Kapitean in age and therefore not restricted to Early Kapitean, as suggested by Hayward (1984). The biostratigraphy presented in King and Thrasher (1996; their Appendix 3) differs again, giving the Ariki Formation a Late Tongaporutuan to mid Kapitean age, and the sequence immediately above a Late Opoitian to Waipipian age. Current data agree with this latter study, with the upper part of the Kapitean Stage, and much of the Opoitian Stage, not identified.

3.4.5 Mangaa-1

Foraminiferal analyses were initially undertaken by Hematite Petroleum (NZ) Ltd. (1970) and later reviewed by Hayward (1985b), in light of the advances made in the understanding of the biostratigraphic ranges and lineages of planktic foraminifera. King and Thrasher (1996) present an age-adjusted version of the Hayward (1985b) report (P. King, pers. comm., 2001), while a study by Waghorn et al. (1996) was initiated as part of a comprehensive review of several wells in the northern Taranaki Basin, with the aim of refining Tongaporutuan to Nukumaruan Stage tops and paleoenvironmental interpretations. In this latter study, all existing faunal slides were re-identified, and combined with new infill samples, greatly increasing sample density (101 cuttings samples and 19 sidewall cores analysed, cf. 33 cuttings samples and 17 sidewall cores analysed in Hayward, 1985b). A comparison of the biostratigraphy presented in King and Thrasher (1996) and that of Waghorn et al. (1996) is given in Table 3.6. This study follows the biostratigraphic scheme of Waghorn et al. (1996; Fig. 3.7), although these authors caution that, because of problems with downhole contamination and prolific reworking of fauna encountered at Mangaa-1, some horizons are not well resolved, and ages may be poorly constrained in places.

Table 3.6: Biostratigraphic schemes for Mangaa-1, after King and Thrasher (1996) and Waghorn et al. (1996) Total depth of well is 3553 m..

King and Thrasher (1996)		Waghorn et al. (1996)	
Depth (m)	**Stage**	**Depth (m)**	**Stage**
0-450	undated	0-450	undated
450-1750	Wn	450-1829/1838	Wn
ca. 1750 –1880	Wp - Wm	1864 – 1907	Wp-Wm
1880-2300	Wp	1907-2414/2423	Wp
2300-2530	l. Wo	2441/2451-2752/2761	Wo
2530-2786	m. Wo	2761/2771-3155/3158	l.Tt-e.Tk
2786-?	e. Tk	3240/3243	e.Tt
3240	e.Tt	3250-3553	Undated Miocene
3250-3553	Undated Miocene		

The differences in placement of New Zealand Stage boundaries between these reports is primarily a function of resolution of sampling intervals. However, Waghorn et al. (1996) identify an unconformity between Opoitian and Early Kapitean strata whereas King and Thrasher (1996) do not. While the earlier report of Hayward (1985b) stated that there were no recognisable time gaps in the Kapitean to Nukumaruan sequence, evidence for this unconformity can be observed in the seismic record, correlating to a bold seismic horizon, and lithologically represented by a marly unit capped by a distinct log break on geophysical wireline logs. Similarly, Strong et al. (1996) indicate that seismic and foraminiferal evidence reveal an unconformity between Kapitean and Opoitian strata at Awatea-1, ~11.5 km to the southwest of Mangaa-1. Seismic correlation clearly links these two horizons (Chapter 5).

3.5 Chronostratigraphic scheme for northern Taranaki Basin.

3.5.1 Composite biostratigraphy

Analysis of the four well sections reviewed in this study, together with the biostratigraphical review of Mangaa-1, confer mixed results in terms of refining the occurrence of stages and the placement of stage boundaries. Where foraminifera are abundant, particularly planktic faunas (deeper well sites), higher resolution sampling can aid greatly in refining stage tops (e.g., Ariki-1, Wainui-1). However, at sites where it is difficult to obtain large numbers of planktic faunas, particularly age-diagnostic species (e.g., Arawa-1, Kora-1), this study has found that higher resolution sampling does little to redefine the previous biostratigraphy. These varied results suggest that the biostratigraphic schemes already in place for other wells in the study area (which have either been age–adjusted from the original report, or have utilised more recently published data) are sufficient to construct an approximate composite biostratigraphic scheme, but could do with some refining should any group be interested. This biostratigraphic scheme can in turn be used as a proxy for a chronostratigraphic framework for northern Taranaki Basin.

Appendix 2 (composite well logs) presents the biostratigraphy for each well section as used in this study. Each of these well sections is linked by way of seismic reflection profiles to other well sections. Numerous seismic units have been identified and mapped across the seismic grid (refer to Chapter 5), and these units are used to correlate stages within the study area. Time-depth conversion has been accomplished using data supplied by Geosphere Exploration Ltd. (Wellington). A composite biostratigraphy has thus been achieved by comparing the age of seismic units at each site (Fig. 3.8a,b).

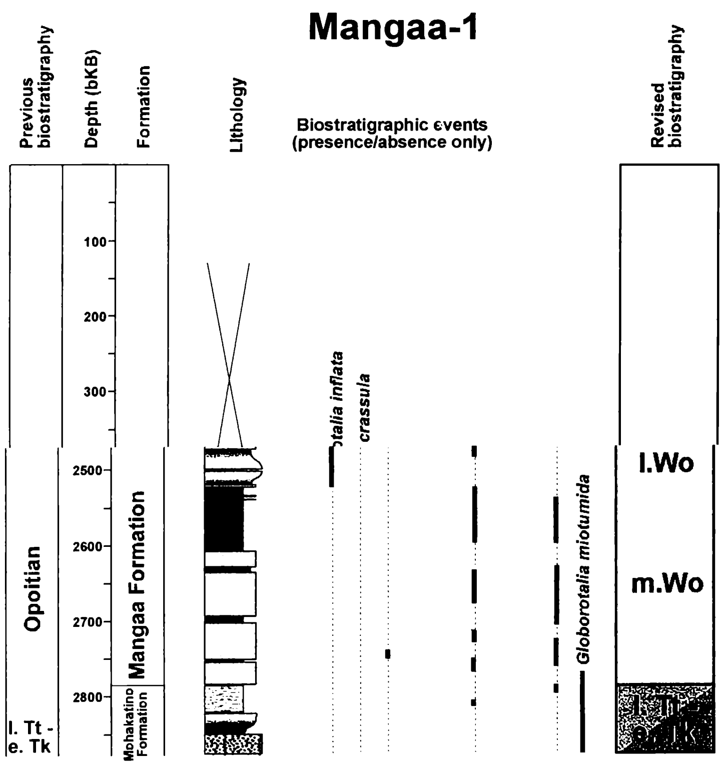


Fig. 3.7: Revised biostratigraphy for Mangaa-1, to 2870 m (after Waghorn et al., 1996). Previous biostratigraphy after Hayward (1985b).

Wainui-1, Arawa-1, and Taimana-1 (Fig. 3.8b) have been treated separately because of their location relative to the cluster of wells in the northern part of the study area (Fig. 3.8a), as seismic units in the southern and western parts of the study area are not able to be correlated to seismic units to the north and east.

Figures 3.8a and b illustrate that overlap between adjacent Plio-Pleistocene stages is common, with no stage being entirely separate from another. However, Taranaki Series strata do not overlap Wanganui Series strata. Much of the inconsistency between well sites can be attributed to the inability to accurately position stage boundaries, or indeed identify some stages (particularly the Mangapanian Stage, and separation of the Nukumaruan Stage from the Castlecliffian Stage) because of the nature of the samples. This is a particular problem at Arawa-1 because of the paucity of faunas, and the inability to accurately constrain the Nukumaruan and Mangapanian Stages. It may also in part be due to unrecognised interpretational errors on seismic reflection profiles.

Nukumaruan to Castlecliffian (2.28-0.33 Ma) sediments dominate the stratigraphic extent of the Giant Foresets Formation in the northern and eastern parts of the study area, at some sites comprising over 90% of the sequence (e.g., Te Kumi-1; see Appendix 6), while to the south and west, Mangapanian to Nukumaruan (2.79-2.28 Ma) sediments predominate. This pattern is primarily related to the progradation of the foreset front past each well site. To the south, the continental margin prograded past Taimana-1 and Arawa-1 during the Late Opoitian to Mangapanian, while the front did not reach the north until the Nukumaruan. Most wells in the northern region record relatively thin Opoitian to Mangapanian sequences, with the exception of Awatea-1 and Mangaa-1. At these two sites, the thick Pliocene-aged, sand-dominated Mangaa Formation underlies the Giant Foresets Formation. Only a thin interval (~160 m) of post-Miocene (Giant Foresets Formation) strata, tentatively dated as Pleistocene (on the basis of single sampled interval; Shell BP Todd, 1975) is present a Turi-1 in the east. This well section sits atop a fault block that underwent uplift and erosion during the Pliocene and Early Pleistocene, evidenced by concretionary cobbles of a lithology similar to the underlying Mount Messenger Formation (Shell BP Todd, 1975).

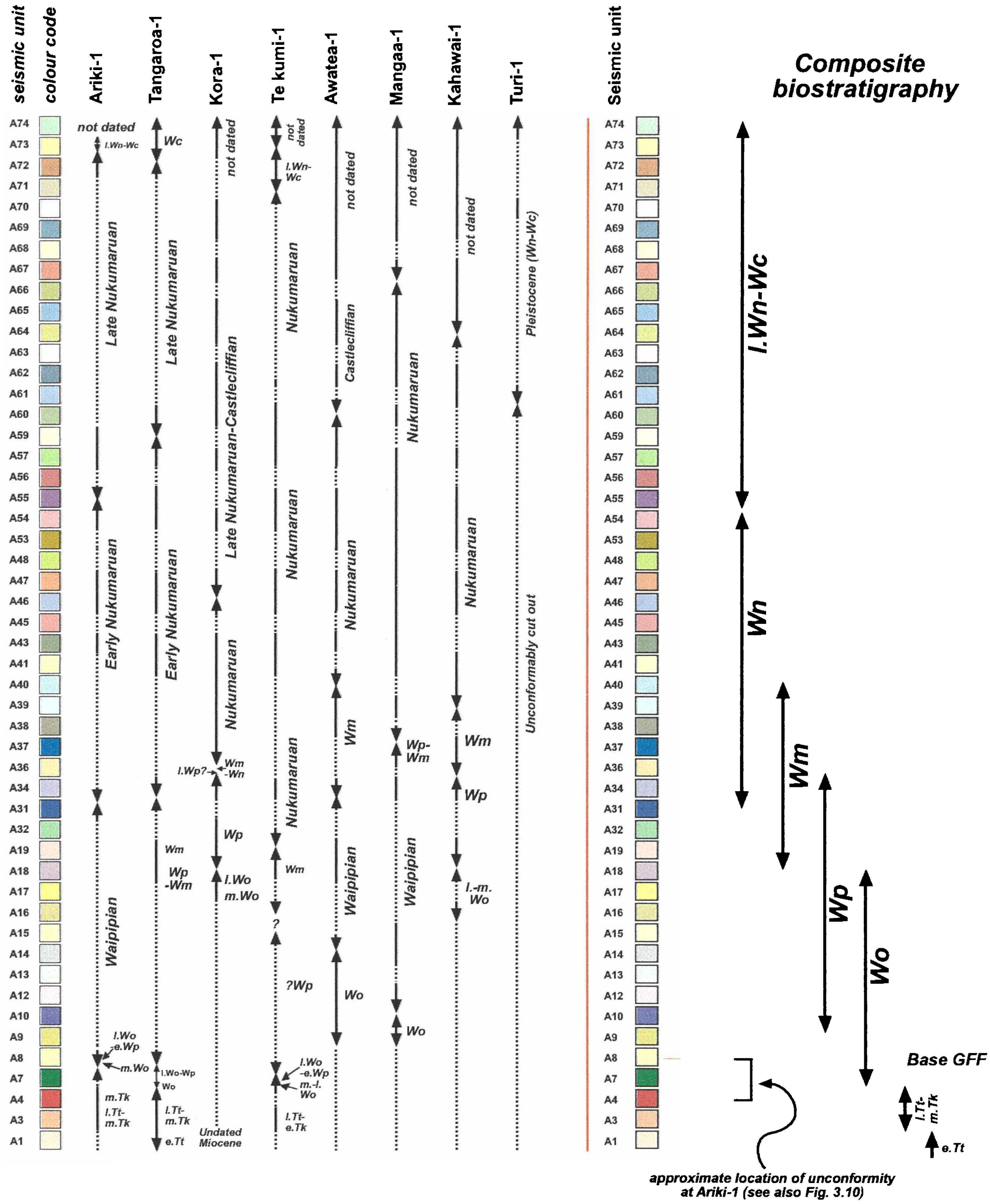


Fig. 3.8a: Composite biostratigraphy (based on New Zealand Stages) for northwestern Western Stable Platform and Northern Graben well sections. For each well section, solid line represents presence of seismic unit, dashed line indicates seismic units not present at that site. Arrows indicate stratigraphic extent of biostratigraphic stages. Refer to Appendix 6 for time-depth correlation of seismic units with the biostratigraphic scheme for individual well sections.

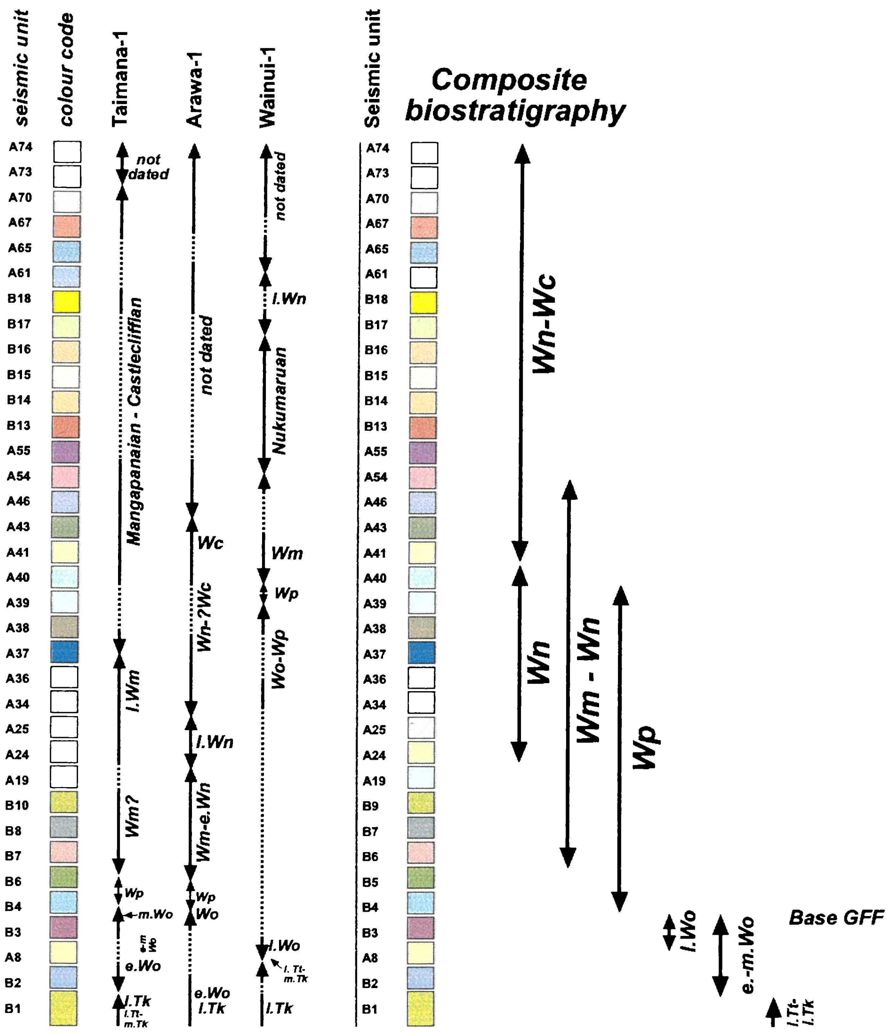


Fig. 3.8b: Composite biostratigraphy (based on New Zealand Stages) for western and southern Western Stable Platform well sections. For each well section, solid line represents presence of seismic unit, dashed line indicates seismic units not present at that site. Arrows indicate stratigraphic extent of biostratigraphic stages. Refer to Appendix 6 for time-depth correlation of seismic units with the biostratigraphic scheme for individual well sections.

At no site within the study area is the Giant Foresets Formation older than mid to Late Opoitian (c. 4.5 Ma or younger), though it is slightly older in southern Taranaki Basin (see section 3.5.4). At some sites however, the base of the formation is considerably younger. At Awatea-1 and Mangaa-1, Opoitian to Waipipian strata unconformably overlie muddy and/or volcaniclastic sediments of the Manganui and Mohakatino Formations, respectively. The Early Pliocene Mangaa Formation is much thicker than equivalent aged sequences at other well sites, and lithologically distinct, comprising thick-bedded and amalgamated sandstone beds that grade up into the siltstone and mudstone beds of the Giant Foresets Formation.

The base of the Giant Foresets Formation also falls within lower Waipipian strata at Arawa-1. The boundary between the Giant Foresets Formation and the underlying (lithologically similar) Manganui Formation appears to be based on the presence of a seismically distinctive horizon and corresponding wireline log break. This same seismic horizon is picked as the base of the Giant Foresets at Taimana-1. At both sites the lower boundary is associated with a thin (few 10's of metres) and highly calcareous siltstone to silty limestone (slightly older at Taimana-1), delineated by a prominent planktic abundance spike relative to the overlying Giant Foresets Formation (refer to Chapter 6). The sequence below this calcareous unit at Taimana-1 was formerly associated with basin floor facies (or bottomsets – see Chapter 5) by Crundwell et al. (1994), but the more recent study by Scott et al. (in prep.), and accepted by this study, show that seismic reflection data display a more progradational geometry (their progradational lobe #3), to at least 1700 m bKB. The base of this progradational sequence occurs within Opoitian strata, and Scott et al. (in prep.) indicate that it was emplaced basinward of the main foreset front. The seismic reflector corresponding to the base of this progradational lobe can be traced to Arawa-1, where it occurs at approximately 1775 m bKB, also within Opoitian strata. While this study is not suggesting that the interval between the base of the Giant Foresets Formation and the Miocene-Pliocene boundary be renamed, it does indicate that the uppermost part of the Manganui Formation at Taimana-1 and Arawa-1, and the Giant Foresets Formation are part of the same progradational system.

The Giant Foresets Formation, as defined in this study, extends upwards to the modern sea floor.

3.6 Development of the Miocene-Pliocene unconformity

Two chronostratigraphic panels were constructed to (a) to highlight the distribution of the Late Miocene to Pleistocene succession in northern Taranaki Basin; (b) illustrate the linkage to southern Taranaki Basin; and (c) to emphasise the nature and extent of unconformity development. The construction of these panels was achieved by plotting the distribution of formations with time for each well section and marking in the extent of unconformities (see Fig. 3.9 for well locations and transects). The cross-section in Figure 3.10 links well sections in the study area, and is oriented approximately southwest to east. The cross-section in Figure 3.11 is oriented north to south, and incorporates information from well sections south of the study area.

Figure 3.10 illustrates the distribution of the Late Miocene-Pleistocene succession across the study area, and highlights the extensive nature of unconformity development across the Western Stable Platform and Northern Graben. As illustrated by Fig. 3.10, the unconformity straddles Miocene to Pliocene strata, and is expressed differently across the study area. On the Western Stable Platform and within the Northern Graben, the unconformity is associated with the Ariki Formation and other marly and calcareous age-equivalent units. To the east, unconformity development is associated with uplift and erosion across the Turi Fault Zone, although at Kahawai-1, it is uncertain as to whether all unrepresented time is associated with erosion, or whether some is incorporated into a thin (c. 9 m) tuffaceous and marly unit at the base of the Giant Foresets Formation. Along the axis of the Northern Graben, Pliocene sediments lie unconformably over extensive areas of the Miocene volcanic massifs.

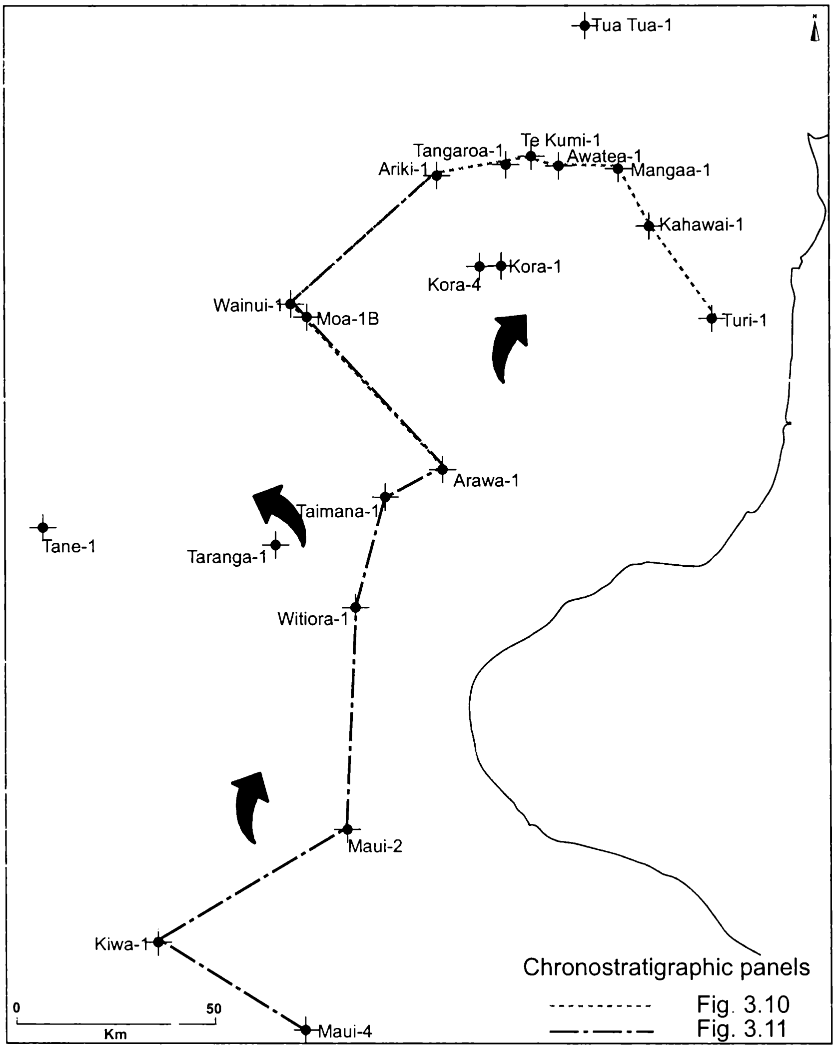


Fig. 3.9: Location of well sections used in Figs. 3.10 and 3.11. Arrows indicate general direction of progradation of the Giant Foresets Formation.

The relationship between the sedimentary succession in northern Taranaki Basin, and the succession in southern Taranaki Basin is illustrated in Fig. 3.11. This figure displays several features, including the limited extent of the Mohakatino and Ariki Formations, and the extensive distribution of the Giant Foresets Formation. It also clearly illustrates unconformity development south of Witiora-1 (note it is unclear whether time is indeed missing at Maui-2). This unconformity development is related to uplift and contemporaneous wave-base erosion during the Late Miocene-Pleistocene, with an increasing age span of missing section southwards a result of increased uplift and tilting in the south (King and Thrasher, 1996). The Miocene-Pliocene succession in the vicinity of Witiora-1, Taimana-1, and Arawa-1 is conformable.

3.6.1 Nature and significance of the Ariki Formation

The Ariki Formation is an important unit in northern Taranaki Basin, marking a period of significant terrigenous sediment starvation that coincides with late stages in the accumulation of the Whangamomona Group/Megasequence. It is a relatively thin and highly condensed unit, reaching a maximum thickness of ~109 m at Ariki-1, and is characterised by an abundant planktic foraminiferal content, forming under more oceanic water masses than the overlying Giant Foresets Formation.

The Ariki Formation is best developed across the Western Stable Platform, an area that has been subject to very little tectonic deformation during the Neogene. It is identified as far southwest as Tane-1 (Shell BP Todd, 1976), and as far north as Tua Tua-1 (Rankin and Barbaresig, 1988). To the south it grades laterally into contemporaneous basin floor mudstone and poorly developed turbidites (King and Thrasher, 1996). Two calcareous to marly units (upper and lower) are identified within the Northern Graben at Awatea-1 and Mangaa-1 (see discussion in Chapters 4 and 5). Lateral continuity between these calcareous units in the Northern Graben, and the Ariki Formation on the Western Stable Platform is questionable, mainly because of post-Miocene faulting (Fig. 3.10). However, it is probable that these units, being of comparative ages to that encompassed by the Ariki Formation, and similarly associated with a Miocene-Pliocene unconformity (lower unit), are related. The Late Miocene and Early Pliocene calcareous units identified at Taimana-1 and Arawa-1 (Fig. 3.11.) may also be related to the Ariki Formation, although no unconformity is associated with these thin units.

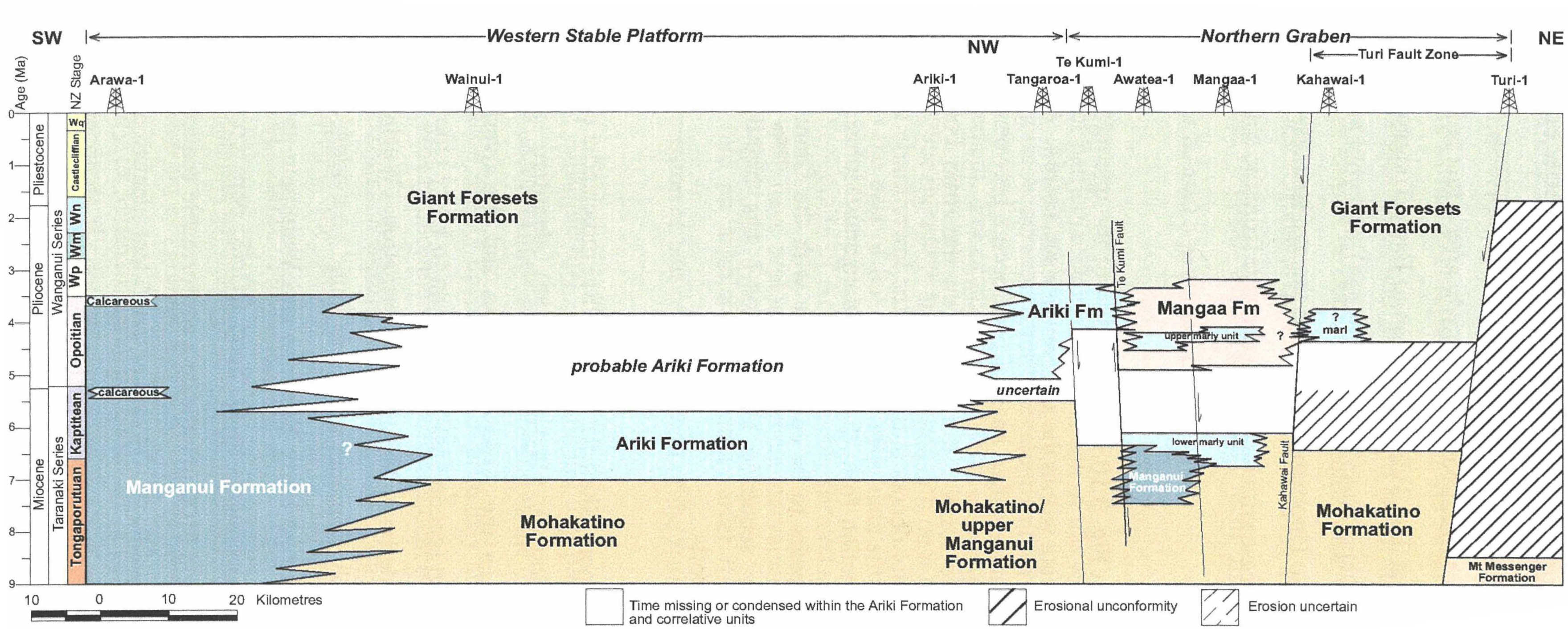


Fig. 3.10: West-east chronostratigraphic panel, northern Taranaki Basin. Note the complex distribution of the Ariki Formation and association to other marly equivalents in the Northern Graben. While the Ariki Formation is only ~109 m at it thickest (Ariki-1), it encompasses a wide time range. It is postulated that on the Western Stable Platform and within the Northern Graben, much of the missing time is condensed within the Ariki Formation or equivalents, and certain stages cannot be identified because of sampling resolution (interval spacing and mixing of samples) and degree of condensation. In these regions, the unconformity is thus better termed a paraconformity. In comparison, the unconformity across the Turi Fault Zone is associated with uplift and erosion, and is often angular in nature. Stage abbreviations; Wq, Haweran; Wn, Nukumaruan; Wm, Mangapanian; Wp, Waipipian.

The period of time encompassed by the Ariki Formation varies between sites, as does the age span of missing or unidentified time. As the development of the Ariki Formation is primarily linked to terrigenous sediment starvation, one would intuitively expect the age of the Ariki Formation to be oldest in the most distal parts of the study area – this is indeed the case, with the base of the formation oldest at Tua Tua-1 (Waiauan to Early Tongaporutuan), younger at Wainui-1 and Ariki-1 (Late Tongaporutuan to Mid Kapitean), and younger still at Tangaroa-1 and Te Kumi-1 (latest Kapitean to Opoitian). Similarly, the age span associated with missing sedimentary section tends to increase with increasing distance from the contemporary landmass (Fig. 3.10).

An important feature noted from Fig. 3.10 is the position of the unconformity relative to the placement of Ariki Formation strata within the sedimentary succession. At Wainui-1, Ariki-1, and Tua Tua-1, the missing or unidentified span of time occurs between the Ariki and Giant Foresets Formations, while at Te Kumi-1 it occurs *below* the Ariki Formation. Within the Northern Graben the unconformity occurs at the top of the lower marly to calcareous interval that separates the Mangaa Formation from the lower Manganui/Mohakatino Formation. At Tane-1 the Ariki Formation appears to be conformable with the Giant Foresets Formation. Evaluation of the sedimentary sequence at Tangaroa-1 has highlighted a number of inconsistencies between well reports.

The original well completion report (Shell BP Todd, 1981) indicated a conformable Neogene sequence, but later re-evaluation by Hayward (1985b) identified an upper Waipipian-Mangapanian time gap (between the Ariki Formation and the overlying Giant Foresets Formation) and a lower time gap (encompassing Kapitean to Early Opoitian Stages) between the Ariki Formation and the underlying Mohakatino Formation. Waghorn et al. (1996) later identified Mangapanian and Early Kapitean strata, but found no evidence of Waipipian strata. King and Thrasher (1996) identified Early Kapitean and Waipipian-aged strata, but did not identify Mangapanian strata. Comments by G. Scott (IGNS; pers. comm., 2001) regarding critical samples between 1860 and 2100 m indicate a conformable sequence to at least 2100 m bKB (lower Opoitian?; see Appendix 2). However, confident identification was unattainable for many of the lower samples because of poor preservation. If an unconformity does exist at this site, it probably occurs near or at the lower boundary of the Ariki Formation (~ 2105 m).

The discussion above, particularly that pertaining to Tangaroa-1, draws attention to the exact nature of the unconformity associated with Ariki Formation development. This study suggests that, at least across the Western Stable Platform and within the Northern Graben, missing sedimentary section is in fact encompassed within the Ariki Formation and equivalent aged marly or highly calcareous units, and just unable to be identified. The fact that sedimentary section of a certain age cannot be documented at some sites, but can at others, is defined as a function of sampling resolution and the homogenous mixing of sample intervals that is inherent with cuttings samples. Thus, where associated with marly and condensed units, the unconformity is better termed a paraconformity.

While identification of particular stages across much of the western region of the study area is considered a sampling issue, paraconformity development is emphasised in places due to local unconformity development. For example, at Te Kumi-1, unconformity development may have been exacerbated by Pliocene uplift and non-deposition on a fault (informally called Te Kumi Fault) immediately to the east of this site. Seismic data similarly suggest that a low dome-shaped paleohigh in the vicinity of Ariki-1, possibly formed as a rollover anticline adjacent to the Northern Graben, may have prolonged sediment starvation in this region. To the east, the paraconformity runs into an angular unconformity over the Turi Fault Zone, a result of significant uplift and erosion of part of the footwall of the fault zone.

Figures 3.10 and 3.11 illustrate that Ariki Formation and paraconformity development were coincident with major unconformity development across the Turi Fault Zone and the Southern Inversion Zone. Paraconformity development is primarily attributed to the limited extent of accumulation of the Whangamomona Group/Megasequence continental margin wedge that prograded northward through Wanganui and King Country Basins during the Tongaporutuan to Early Opoitian, involving the Mount Messenger, Urenui, Kiore, and Matemateaonga Formations (Kamp et al., 2002). The condensed Ariki Formation in northern Taranaki Basin reflects the focus of contemporary terrigenous sedimentation to the east in King Country Basin and along the eastern margin of Taranaki Basin, as well as possible oceanographic changes that lead to more oceanicity in northern Taranaki Basin for a period. In contrast, unconformity development across the Turi Fault Zone is a result of extensional tectonics, while to the south (Southern Inversion Zone), it is a result of compressional tectonics, both a physical manifestation of plate boundary development.

3.7 Conclusions

- Revision of the biostratigraphy of Arawa-1, Ariki-1, Kora-1, and Wainui-1, and a review of the latest version of the biostratigraphy of Mangaa-1, has revealed mixed results in terms of the effectiveness of more detailed biostratigraphic evaluation of well sections. Resolution of biostratigraphic stages have been increased dramatically at sites that have remained at deeper water depths, and which contain abundant planktic foraminifera, but more detailed sampling does little to refine biostratigraphic boundaries at sites that shallowed earlier due to progradation of the continental margin and where planktic foraminifera are not common.
- The lower boundary of the Giant Foresets Formation is often associated with an unconformity that encompasses the Miocene-Pliocene boundary. The unconformity embraces an increasingly older and longer-lived time span to the north, while the base of the Giant Foresets Formation also becomes increasing younger to the north.
- While the age of the base of the Giant Foresets Formations is diachronous, it generally falls within the mid to Late Opoitian interval. At Awatea-1 and Mangaa-1, the base of the formation is Waipipian in age. At these two sites, the Giant Foresets Formation is underlain by the Opoitian-Waipipian aged Mangaa Formation, which grades conformably into the Giant Foresets Formation. At Arawa-1, and Taimana-1, Manganui Formation sediments grade conformably into Giant Foresets Formation sediments. At Taimana-1, a mid Opoitian age for the base of the Giant Foresets Formation is inferred, while at Arawa-1, a Waipipian age is interpreted (seismic evidence suggests that the boundary could be placed lower, within the Opoitian Stage).
- The Miocene-Pliocene unconformity is expressed differently across northern Taranaki Basin. In the northwestern regions of the Western Stable Platform and within the Northern Graben, the unconformity is associated with a highly condensed unit (Ariki Formation) or equivalent aged marl and marly siltstone, and is associated with terrigenous sediment starvation, although may be modified in places by local topography. Inability to recognise certain intervals of time above or below the Ariki Formation is attributed to sampling resolution of well cuttings, and thus the unconformity associated with these condensed intervals is better termed a paraconformity. South of the study area (southern Taranaki Basin) and east across the Turi Fault Zone, unconformity development is associated with uplift and erosion of

the footwall of the fault zone. Along the central axis of the Northern Graben, Pliocene sediment rests unconformably over Miocene volcanic massifs.

Chapter 4: Geophysical wireline log characteristics of late Neogene sediments, offshore northern Taranaki Basin

Chapter 4: Geophysical wireline log characterisation of late Neogene sediments, offshore northern Taranaki Basin

4.1 Introduction

Wireline geophysical well measurements (or more commonly, wireline logs, well logs, or electric logs) are defined as the continuous recording of a geophysical parameter along a borehole (Rider, 1986). Wireline logs are one of the principal tools used for studying subsurface rocks, with the primary objectives of discovering (a) which formations contain hydrocarbons, and (b) how much can be recovered. They are used extensively in petroleum exploration to image the physical, electrical and radioactive properties of formations, and to identify productive zones, determine depth and thickness of these zones, to distinguish between oil, gas, or water in a reservoir, and to estimate hydrocarbon reserves (Cant, 1992; Asquith, 1982). Wireline geophysical measurements have also proven to be useful in the interpretation and downhole mapping of paleoenvironments.

This chapter applies wireline logging interpretative methods to the late Neogene sedimentary succession intersected in wells drilled within the study area. Wireline log shapes or motifs (in conjunction with well cuttings and sidewall cores) are used extensively as a means of describing the subsurface characteristics of the Neogene formations (section 4.3) and for determining wireline facies (section 4.4) and depositional paleoenvironments (section 4.5). Where possible, outcrop observations are used to substantiate subcrop interpretations. A brief examination of the stratigraphic relationships between Taranaki Basin and Wanganui Basin successions is also undertaken in section 4.3. The recognition of sedimentary cycles in Wanganui Basin sequences based on wellcore observations and associated wireline log character have been useful in establishing wireline log motifs and hence the identification of cyclical packages through the Giant Foresets Formation (section 4.6). Graphic logs for each well discussed in this study are given in Appendix 2; these columns also display interpreted lithology, stratigraphic nomenclature, and biostratigraphic age. Note that these well logs typically display the late Neogene section only, and do not continue to the base of the well.

4.1.1 A brief overview of geophysical wireline logging

Since the first electrical log was recorded in France, in 1927 (Doveton, 1994), they have become an indispensable and invaluable tool in the petroleum exploration industry. Various properties and parameters of subcrop, such as lithology, mineral content, chemical composition, porosity, permeability, and fluid content can be analysed. Wireline logs can also be used to support stratigraphic correlation in the subsurface, determine the dip and strike of a unit, and assist in identifying tectonic structures (Zimmerle, 1995; Cant, 1992). Mapping of subsurface facies relationships, essential for the identification of future drilling locations, is primarily achieved through recognition of characteristic wireline signatures or 'motifs', which can often be correlated from well to well.

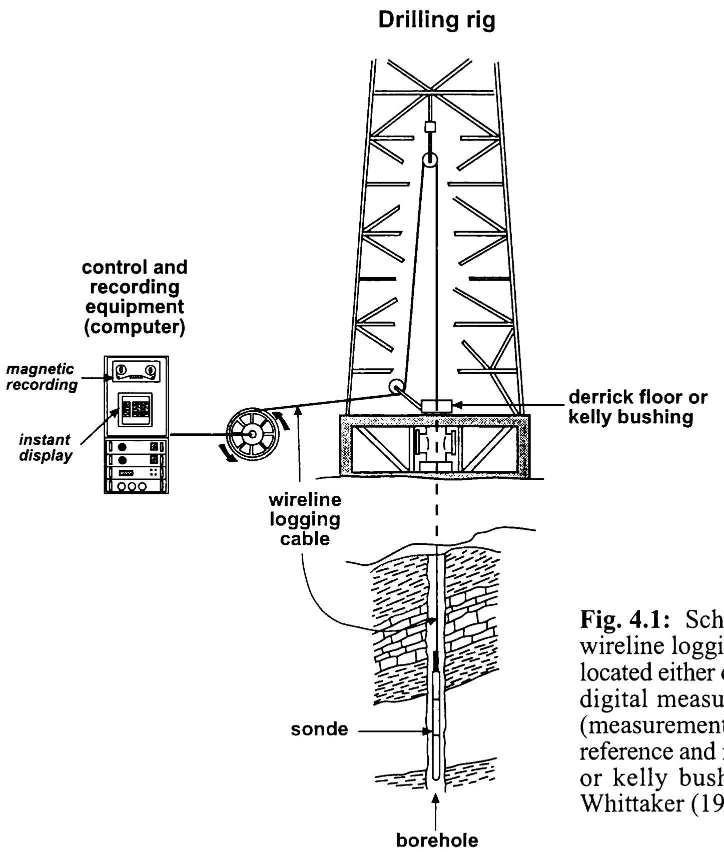


Fig. 4.1: Schematic diagram of a modern geophysical wireline logging set-up in a borehole. The well may be located either onshore or offshore. The computer records digital measurements transmitted by the logging tool (measurements also stored on magnetic tape for future reference and interpretation). Zero datum is derrick floor or kelly bushing. Adapted from Rider (1986) and Whittaker (1998).

Wireline logs are acquired by a petroleum exploration company as the well is drilled. Before the well is classified as an oil or gas producer, or plugged and abandoned as a dry hole, a measuring instrument (or sonde), connected to a cable (wireline) is lowered to the bottom of the borehole (Fig. 4.1). This sonde is slowly winched back to the surface, measuring various parameters of the rock section, such as its resistivity to electrical current, the natural radiation emitted, or transmittance of induced vibrational waves, as a function of depth. Logs are recorded digitally in the computer that is connected to the wireline cable (Rider, 1986). Depth to a particular bed can be established by subtracting the absolute depth in the well, from the surveyed elevation of Kelling Bushing (KB) on the drilling platform (or derrick floor). In this study, all well depths, unless otherwise specified, are referred to in driller depths; i.e., in metres below Kelly Bushing (m bKB).

This study utilises several types of wireline logs, principally gamma-ray (GR), spontaneous potential (SP), and resistivity (shallow and deep) logs. Density, sonic, and caliper logs, and occasionally dipmeter logs, were also utilised. Table 4.1 summarises the types of logs discussed in this chapter, and the properties measured with each. Appendix 4a presents a more thorough discourse on each of the log types, including the parameters measured, common log shapes, and inherent problems. Definitions of log abbreviations are listed in Appendix 4b.

Table 4.1. Log types, properties measured, and geologic uses (from Cant, 1992).

Log	Property measured	Units	Geologic uses
Gamma-ray	Natural radioactivity – related to K, Th, U	API units	Lithology (shaliness), correlation, curve shape analysis
Resistivity	Resistance to electrical current	Ohm-metres	Identification of coals, bentonites, fluid evaluation
Spontaneous Potential	Natural electrical potential (cf. drilling mud)	Millivolts	Lithology (in some cases), correlation, curve shape analysis, identification of porous zones
Density	Bulk density (electron density)	gm/cm^3	Identification of some lithologies such as anhydrite. halite, non-porous carbonates
Neutron	Concentrations of hydrogen (water and hydrocarbons) in pores	Percent porosity	Identification of porous zones, crossplots with sonic, density logs for empirical separation of lithologies
Sonic	Velocity of compressional sound wave	Microseconds/metre	Identification of porous zones, coal, tightly cemented zones
Caliper	Size of hole	Centimetres	Evaluate hole conditions and reliability of other logs
Dipmeter	Orientation of dipping surfaces by resistivity changes	Degrees (and direction)	Structural analysis, stratigraphic analysis

4.2 Data acquisition and management

Geophysical wireline logs used in this study were obtained from Crown Minerals Group, Ministry of Economic Development (MED) Wellington, in LAS format, either on CD or Zip disc. These had been previously converted from binary format files (*.LIS or Log Industry Standard) to ASCII format (*.LAS, or Log ASCII Format) using the program PETROLOG (Crocker Data Processing, Australia). Files were opened in Excel, converted to text documents, with individual logs then graphed using the Macintosh™ logging program, AppleCORE. These were subsequently saved as *.PICT files, and then opened and edited in Macromedia®

Freehand™ drawing package. Occasionally when required, missing sections of incomplete files were scanned or digitised and incorporated into the composite well log in Freehand™.

4.2.1 Interpretational limitations

Several rock attributes, or combination of attributes (Cant, 1992), may result in the interpretative procedure being somewhat subjective. These include:

1. unusual minerals in the rock;
2. anomalously high or low porosity;
3. thinly interbedded lithologies;
4. poor hole condition
5. poor log quality

Because different rock or pore fluid properties affect wireline logs in different ways (see Appendix 4a), it is important to integrate all available wireline logs. However, interpretation of subsurface lithologies entirely from logs, without any other data, is imprecise; lithologic interpretation reconciling core and log data, particularly where lithologies are known in general, are made with much more confidence (Cant, 1992; Lovell et al., 1998). This study therefore utilises all available information in any interpretations made.

Gamma-ray logs are utilised extensively in this study for facies interpretation. Gamma-ray logs respond to natural radiation in sediments – this radiation tends to increase with increasing clay content (becomes 'dirtier'), and decrease with increasing 'clean' sand. These logs therefore are often used as a proxy for grain size. It is, however, important to understand that GR log shapes do not always correlate faithfully with grain size variations (e.g., Rider, 1990). Instead, GR log response is determined by both a *textural* (finer grained rocks tend to be more radioactive because of clay content) and a *compositional* (dependant on the minerals comprising the sediment) factor. While no detailed studies on the mineralogical composition of the Giant Foresets Formation have been undertaken, lithological logs indicate that the formation is quartz dominant, though in places variously mica-rich. Micas are inherently more radioactive, and may result in a slightly suppressed GR log through these intervals. Calibration against cuttings samples and integration with other logs is necessary for correct interpretation of logs.

Additional problems arise due to the limited number of wells available in the study area, and the distance between many wells. These distances are as great as ~52 km (distance between Arawa-1 and Wainui-1), with no well being closer to the next than ~7 km. This makes correlation difficult at times and it is only with paleontological data supported by seismic reflector correlations that plausible relationships can be inferred between strata of relevant ages. It is unfortunate that in the drilling environment it is not deemed necessary to record or collect data from the upper part of a borehole – in the offshore Taranaki Basin the upper 400-500 m is often cased. This means that, particularly in the more northwestern parts of the field area, much of the shelfal part of a sequence (topsets) is missing from the database.

4.3 Characterisation of late Neogene sediments

Geophysical wireline logs are an important interpretative tool, necessary for achieving subsurface correlation between well sections. Characteristic logs shifts, trends, and peaks can be used to identify formations and unconformities. A primary axiom of sedimentary geology is that the present is the key to the past. However, in this section, the past is the key to the present, for it is only by understanding the stratigraphy, architecture, geometry, and where possible, wireline response, of the formations cropping out along the northern Taranaki coastline that valid predictions can be made for analogous formations in the subcrop. This section therefore not only presents a review of the gross wireline characteristics of the Giant Foresets Formation, but also of the preceding Middle to Late Miocene formations, which both underlie the Giant Foresets Formation offshore, and/or crop out onshore.

4.3.1 Manganui Formation

The Manganui Formation is a mud-dominated interval, and this is reflected in the relatively featureless wireline signatures characteristically associated with this formation. King and Thrasher (1996) note that this formation displays a broadly similar signature basin-wide, incorporating a distinctive arcuate profile on GR logs, with higher GR counts at the top and bottom of the succession, though this characteristic appears to be more evident in well sections in southern Taranaki Basin and those onshore. In northern offshore Taranaki Basin, the GR log is distinctly less arcuate, and instead displays a more linear profile (Fig. 4.2).

The contact between sediments of the underlying Taimana Formation and those of the overlying Mohakatino Formation may be gradual but distinct, to pronounced with a widespread GR break (uphole increase), to diffuse and diachronous in offshore wells (King and Thrasher, 1996).

The widespread Manganui Formation immediately precedes the Giant Foresets Formation at Arawa-1 and Taimana-1, and is encountered beneath the volcanic massif on which Kora-1 has been drilled. Sporadic coarser units are scattered throughout the formation, and these are often indicated by marked deflections (lower values) on GR logs. The coarser units are either individual beds, or may occur as a succession of crudely coarsening-upward cycles. At both Arawa-1 and Taimana-1, the uppermost 200-300 m of the Manganui Formation is noticeably different, with lower GR values, considerably lower resistivity values, and decreasing sonic velocity uphole (Fig. 4.2; see also Appendix 2). At Arawa-1, this trend can be correlated with textural observations, which illustrate a positive relationship between increasing sand percentage, and decreasing GR values uphole. At both these sites, this upper sandy interval is age equivalent to the upper Mangaa Formation sands (4.3.6), and may or may not be related to them. Occasional spikes on the GR log (low API values), mirrored by resistivity and sonic spikes suggest the presence of cemented or condensed horizons, or possibly re-worked shell lags.

4.3.2 Mohakatino Formation

The Mohakatino Formation is clearly defined by its geophysical wireline characteristics. The base of the formation, and also of Miocene volcanic massifs, is most often delineated by a distinctive deflection (uphole decrease) in the GR log, with a corresponding uphole increase demarcating the top of the formation (Fig. 4.3a and b). The basal deflection invariably occurs within Lillburnian-aged strata (implying a Lillburnian age for the onset of volcanism), and can be widely correlated between offshore wells (King and Thrasher, 1996), making this horizon an important Miocene marker in Taranaki Basin.

The Mohakatino Formation is characterised by a low and often linear to blocky GR log signature, even at sites where volcaniclastic lithologies are not recorded in cuttings (Bergman et al., 1990), and displays few grain-size trends. SP logs are often featureless throughout this interval (e.g., Ariki-1 and Kahawai-1, Appendix 2), but may also imitate GR logs.

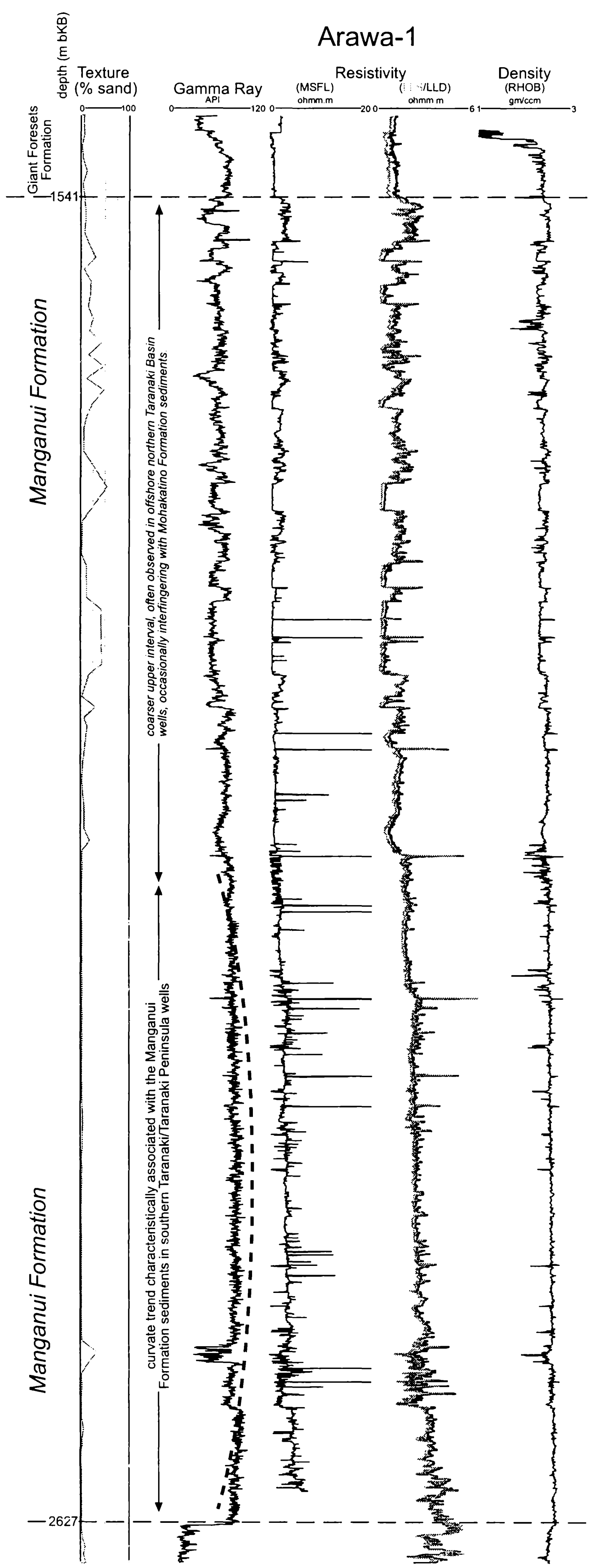
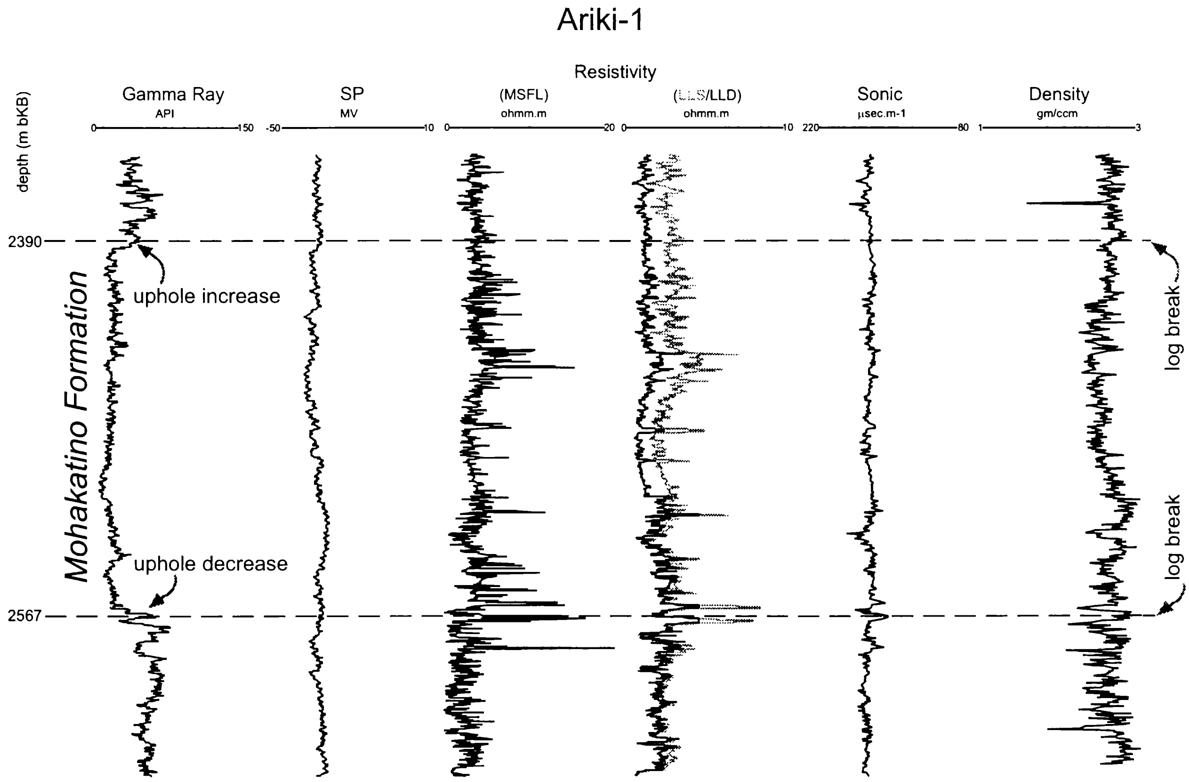


Fig. 4.2: Typical wireline characteristics of the Manganui Formation (northern Taranaki Basin). Example is from Arawa-1. The broadly arcuate profile noted by King and Thrasher (1996) is only evident in the lower part of the formation (dashed line). An overall coarsening-upward trend is supported by the associated textural curve, and which is commonly observed in the top part of the Manganui Formation in offshore northern Taranaki Basin wells. In some wells (see, e.g., Fig. 4.5), this interval is volcaniclastic, and is incorporated into the Mohakatino/ upper Manganui Formation.



(a) Mohakatino Formation

Fig. 4.3: Typical wireline characteristics of volcaniclastic sediments. **(a)** Mohakatino Formation, Ariki-1. Note the GR shift to the left (uphole decrease) at the base of the formation, and back to the right (uphole increase) at the top of the formation; **(b - facing page)** volcanic massif (Kora-1). Wireline characteristics are similar to (a), although all logs display more distinctive kicks at the top and bottom of the volcanic interval.

Resistivity logs invariably become more erratic within the Mohakatino Formation, with higher values being obtained than in under- or overlying formations: offset between shallow (LLS/SFLU) and deep (LLD/ILD) logs also suggest coarser lithologies (e.g., Fig. 4.3a). Sonic (velocity) logs display variable trends between well sections, and may either exhibit slightly higher or slightly lower velocities. Bulk density logs again may be slightly higher, but do not appear to show features that are diagnostic of the Mohakatino Formation only. Overall, the predominantly blocky and often very smooth nature of the GR log suggests that massive (volcaniclastic) sandstone beds comprise much of the formation. This log motif is similar to that interpreted by Salimullah and Stow (1992), who show that this volcaniclastic facies is typical of deep-water slope and basinal environments where volcanism has contributed to the sedimentary input.

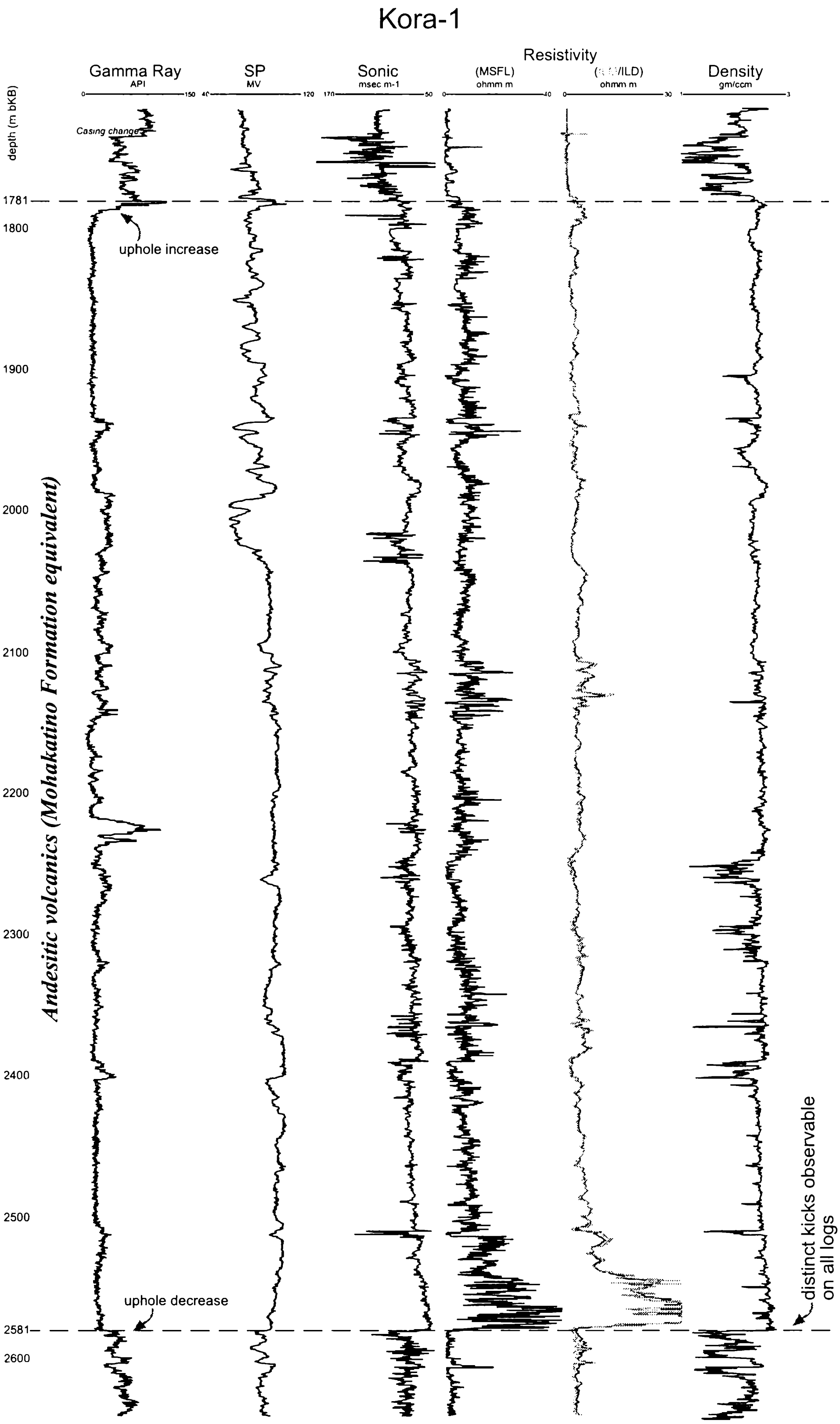


Fig. 4.3(b): Miocene volcanics (Kora volcanic massif).

Lithologic comments noted in well reports suggest that sediment texture is often not as coarse as one might expect from the response illustrated by GR logs. This is exemplified by textural trends at Arawa-1 (see Appendix 2), with the sand fraction (> 63 μm) generally less than 25% and often less than 10%. Low GR values may be attributed to the type of clays present in the formation. Basaltic-andesitic composition rocks of the type that comprise the Mohakatino Formation (Bergman et al., 1990, 1992; Smale, 1992) ultimately degrade to smectitic and montmorillonite-type clays (Leslie et al., 1993; R. Briggs, pers. comm., 2002). Smectite was in fact noted as one of the constituent clay types of the Ngarupupu Formation (a coastal equivalent of the lower part of the Mohakatino Formation; Nodder et al., 1990b). These clays, compared to illitic clays, are non-radioactive, and would therefore decrease expected GR values. The high resistivity and density values also suggest that the clays are not radioactive, and that the formation therefore has unusually low gamma ray values.

4.3.2(i) Mohakatino-Manganui Formation

The Mohakatino Formation varies in thickness between well sites, and the wireline motif differs dependent on one, or a combination of, three factors:

1. distance from source;
2. waning of volcanism;
3. mechanism of distribution (i.e., aerial fall or subaqueous flow).

Waning of volcanism upward within the formation is particularly apparent in well sections proximal to fading eruptive centers, such as Ariki-1 and Tangaroa-1, where interfingering of volcaniclastic sediments with the muddier grain size and siliciclastic composition of the Manganui Formation is evident. In such cases, the dominantly volcaniclastic Mohakatino Formation cannot be separated from the dominantly argillaceous Manganui Formation, and no distinction is made between the two. Log motifs through this interval tend to be characterised by thinner bedded units that display fining- ('dirtying') upward trends (suggestive of a turbiditic mechanism of deposition). These intervals also display an overall fining-upward trend, probably reflecting a decline in volcanic activity.

Outcrop studies utilising a GR scintillometer (Bergman et al., 1990), demonstrate that intervals within the highly volcaniclastic Mohakatino Formation strata (Ngarupupu Formation) exposed onshore behave in much the same way as that described for offshore wells. While such logs cannot be directly compared to offshore logs (units are in counts per second or cps, rather than

API units), it nevertheless reproduces the log response observed in borehole logs. Gamma-ray values are very low compared to other clastic lithologies in the vicinity, and individual beds are slightly blocky to fining-upward (Fig. 4.4), indicating similar mechanisms of emplacement. Nodder (1987) and Nodder et al. (1990b) interpret the Ngarupupu Formation sediments as having been deposited on volcaniclastic aprons distal from the volcanic edifices of the Mohakatino Volcanic Complex.

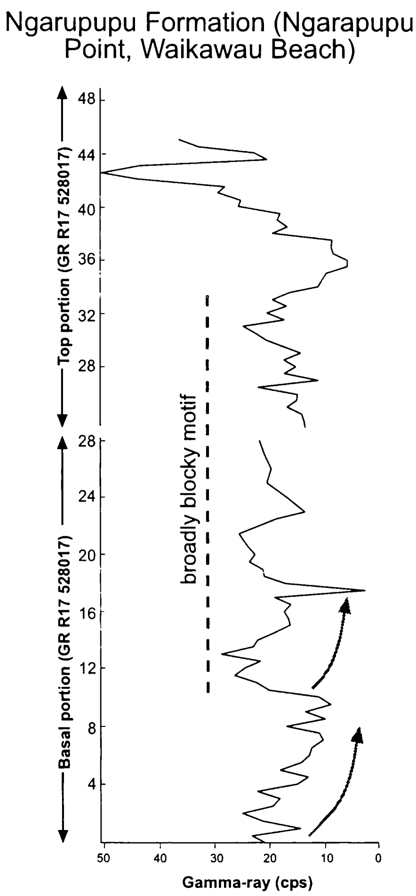


Fig. 4.4: Gamma-ray log characteristics of basal Mohakatino Formation equivalents (Ngarupupu Formation of Nodder, 1987). From Bergren et al. (1990). Height in metres. While still broadly blocky in nature, presence of fining- and coarsening-upward units, similar to those observed in the Manganui/Mohakatino Formation log motif offshore, reflects the distal location of the site onshore (Waikawau Beach) from source.

4.3.2(ii) Lateral variation

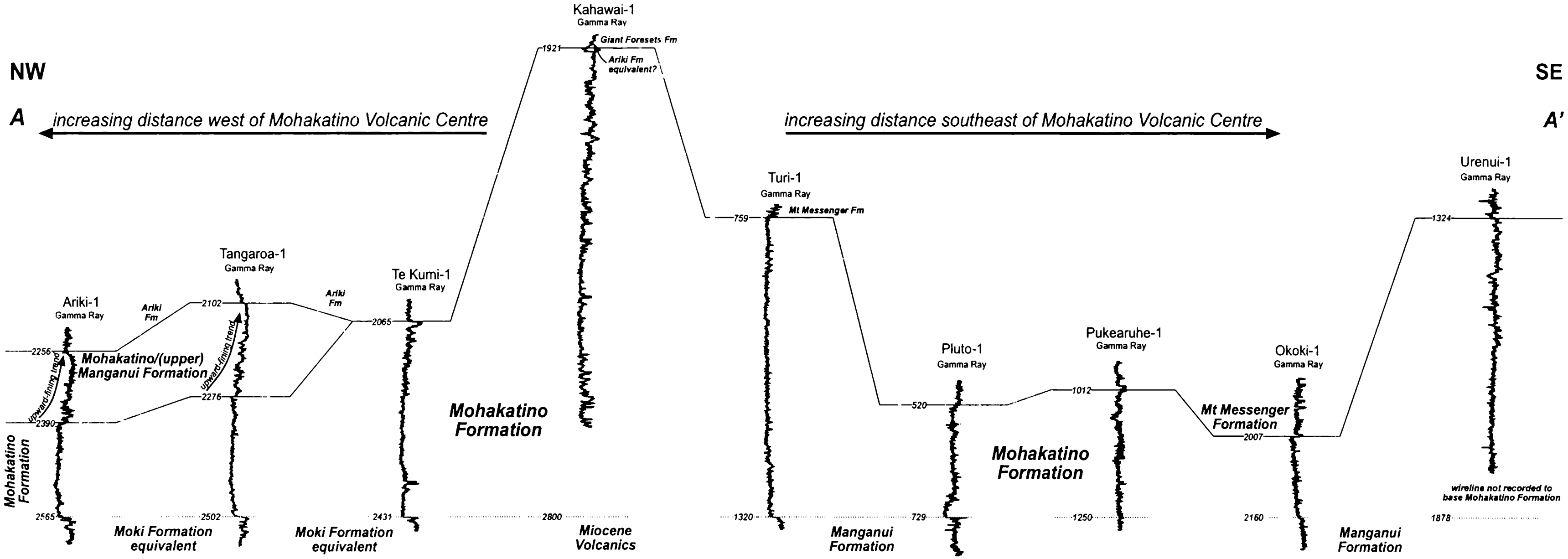
Individual bed thicknesses of outcropping Mohakatino Formation strata have been measured at between a few mm to up to 20 m for amalgamated beds (Nodder, 1987; Hansen, 1996; see also King et al., 1993). Gamma-ray logs obtained at sites distal to volcanic edifices (e.g., Wainui-1, Taimana-1, Tane-1, Mokau-1, and Pluto-1), become more serrate, mimicking the thinner-bedded nature of strata cropping out on the coast (Fig. 4.5). Resistivity logs are much less erratic than at sites proximal to volcanic edifices. The closer particular wells are sited to a volcanic complex, the blockier and more amalgamated the beds become, and the more erratic are the resistivity values, reflecting a highly variable clay content. An uphole decrease in volcaniclastic detritus is reflected by diminishing bed thickness and a commensurate amplification in the degree of serration of the GR log, uphole increasing GR values, and less erratic resistivity. This suggests waning of local volcanism, with the muddy sediments of the Manganui Formation once again dominating. In parts, the Mohakatino Formation looks remarkably similar to the lower part of the Mount Messenger Formation, which points towards similar mechanisms of emplacement, or mild volcaniclastic overprinting of a dominantly terrigenous clastic system, particularly along the eastern margin of the basin.

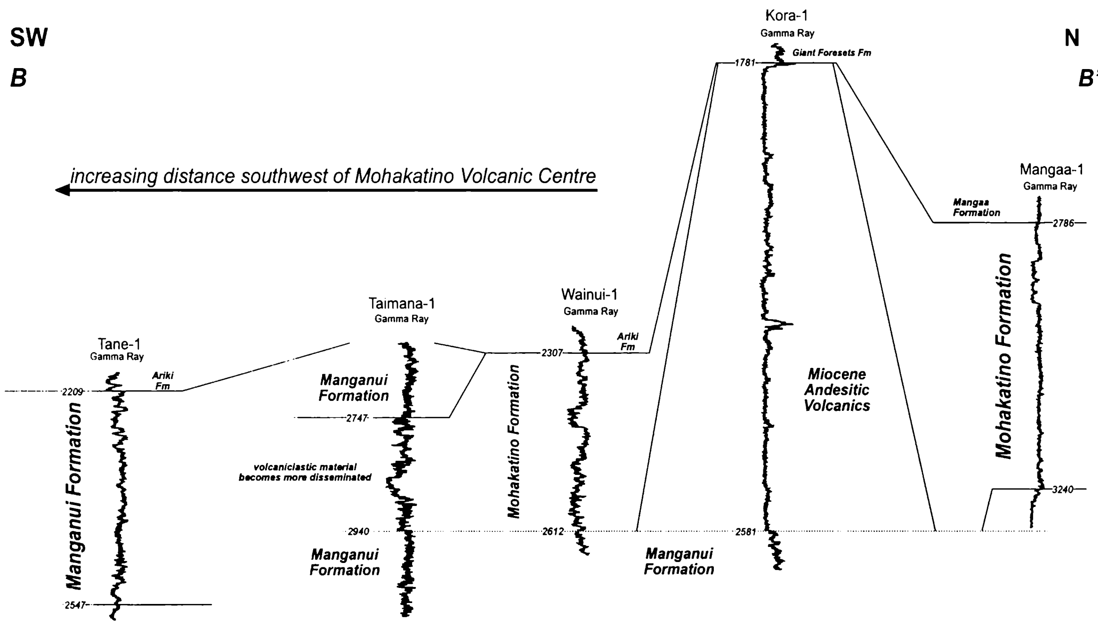
4.3.2(iii) Volcanic massifs

Several wells intersect Miocene volcanic massifs, including Kora-1 to 5, Te Kumi-1, and Mangaa-1. At these sites, undifferentiated volcanics display characteristics similar to the Mohakatino Formation, although there is often a stronger kick at the base and top of the interval than displayed by Mohakatino Formation sediments, and resistivity logs are less erratic (Fig. 4.3b). High degrees of induration or compaction of volcanic massifs is indicated by sonic and density logs, which at Kora-1 and Kora-4 are displayed by a sharp increase at the base of the volcanics, and a decrease at the top.

4.3.3 Mount Messenger Formation

The Mount Messenger Formation is not described in wells within the Northern Graben or Western Stable Platform, although it has been identified in a number of the more coastal sites (e.g., Turi-1, Awakino-1, Okoki-1, and Pukearuhe-1). At several of these sites (e.g., Turi-1, Pluto-1), this formation is present immediately beneath Pleistocene-Recent sediments of the Giant Foresets Formation, the boundary being a significant unconformity. The Mount Messenger Formation has been closely investigated (e.g., King et al., 1993, 1994; Browne et al., 1996; Browne and Slatt, 1997; Browne et al., 2000) in terms of its depositional environment and the mechanisms of emplacement of the component beds; the wireline geophysical characteristics of (parts of) the formation have also been examined (e.g., Jordan et al., 1994; King et al., 1994; Coleman et al., 2000). The outcrop occurrences of the Mount Messenger Formation are an important analogue for subsurface units because of the detail acquired about beds, bed boundaries, and lateral variations in thickness that can be estimated from coastal occurrences.





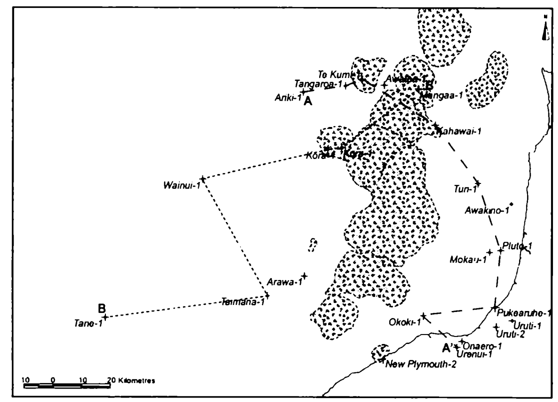


Fig. 4.5: Gamma-ray wireline motif for the Mohakatino Formation, showing lateral variation to the south and east of the Miocene volcanic centre. Note the upward-thinning nature of the Mohakatino/Manganui Formation (Ariki-1 and Tangaroa-1).

In well sections, well-developed massive sandstone beds interpreted as stacked basin floor fan cycles in outcrop are present at the base of the formation (lower Mount Messenger Formation; King et al., 1993). These are best expressed by SP and resistivity logs as a series of tabular or blocky packages with sharp lower and upper contacts, each of which display a broad fining-upwards character (Fig. 4.6). This wireline log motif, in a basinal setting, is indicative of thick bedded and amalgamated units characteristic of a succession of fan lobes sheets or fan channel deposits (King et al., 1993; King and Thrasher, 1996). In outcrop, individual sandstone beds are up to 4-5 m thick, while amalgamated packages can obtain thicknesses of 30-40 metres, and display the sharp lower contacts that are evident on wireline logs (Fig. 4.7). Thick intervals of thinner-beds are prevalent above the basal amalgamated fan deposits (see Fig. 4.7, lower photo); interbedded siltstone, mudstone, and fine-grained sandstone occur in repeated, smaller scale coarsening-upward cycles.

The upper Mount Messenger Formation interval, interpreted from outcrop characteristics as channel levee and overbank deposits in a slope fan complex (King et al., 1994; Browne and Slatt, 2002; Fig. 4.8), is characterised on wireline logs by a stacked succession of upward-fining or upward coarsening cyclothemic packages. In outcrop, the thin-bedded (centimetre-decimetre thick) sandstone beds are finer grained and better sorted than sandstone of the basin floor fans, show well developed Ta-Tc Bouma divisions (Browne and Slatt, 2002), and grade upward into decimetre-thick mudstone dominated (Td and Te) intervals.

Gamma-ray log profiles recorded by King et al. (1994), and a composite log by Diridoni (1994, in King et al., 1994) using a hand-held multi-channel scintillometer (Fig. 4.9), mimic the log pattern displayed in the subcrop setting. Thick-bedded intervals generally exhibit a lower gamma-ray response than the thin-bedded intervals, but overall the formation (both in outcrop and subcrop) exhibits a general fining-upward trend, with up-sequence shallowing indicating slope fan progradation.

Continuity of beds is an important criterion in basin studies, particularly when trying to understand and characterise potential hydrocarbon reservoirs. Slatt et al. (1992) demonstrated that ambiguity occurred between multiple logs when discontinuous strata were present, even where logs were closely spaced. The study concluded that, based on the average spacing of exploration wells, *sequences* of sheet-like, laterally continuous beds could be readily correlated on GR logs, although individual beds could probably be neither distinguished nor correlated.

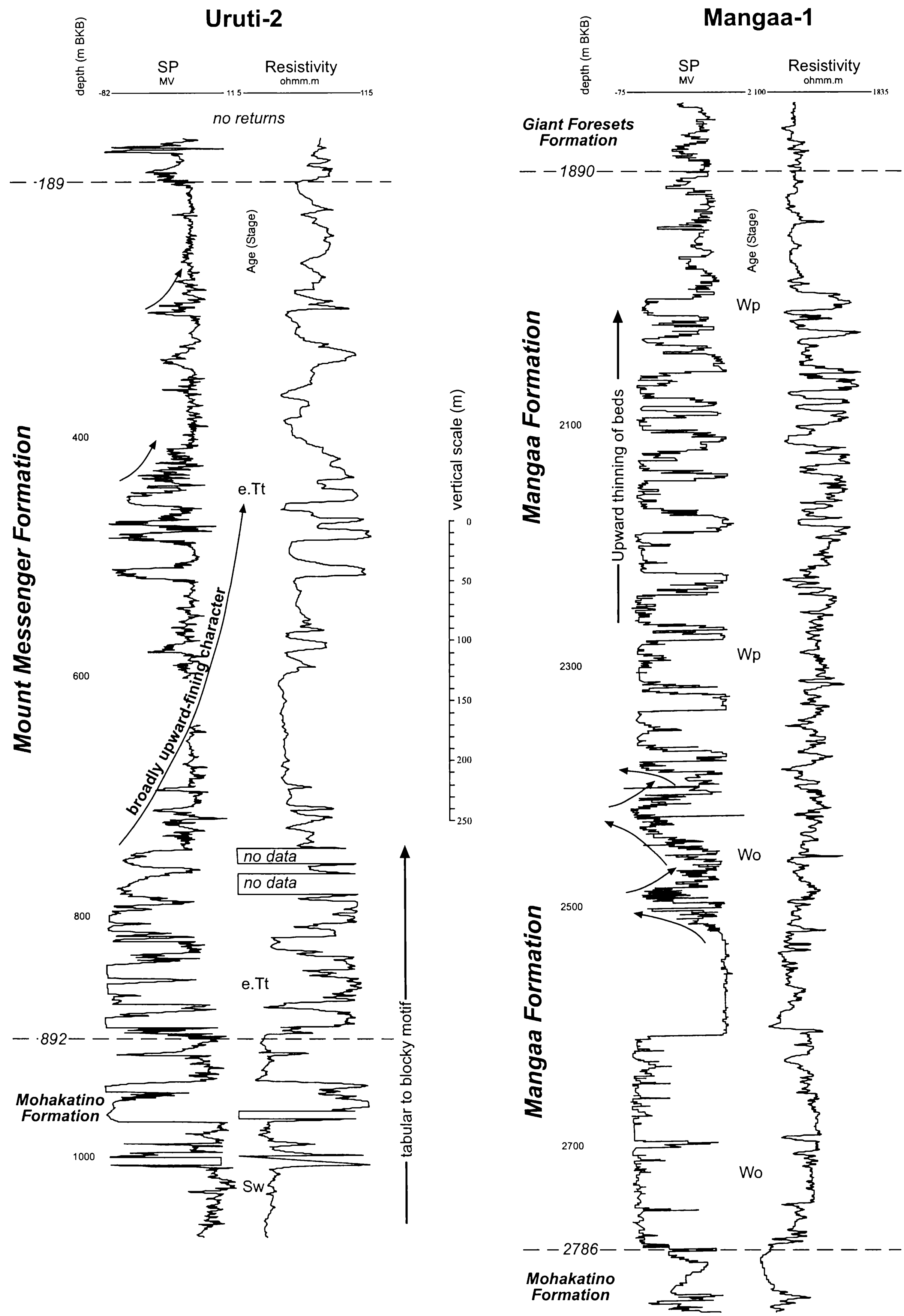


Fig. 4.6: Characteristic SP and resistivity logs, Mount Messenger Formation (Uruti-2, left), and Mangaa Formation (Mangaa-1, right). Note the similarity in stacking pattern, and comparable thicknesses of amalgamated, blocky to tabular sandstone units. New Zealand Stage abbreviations: Sw, Waiauan; Tt, Tongaporutuan; Wo, Opoitian; Wp, Waipipian.

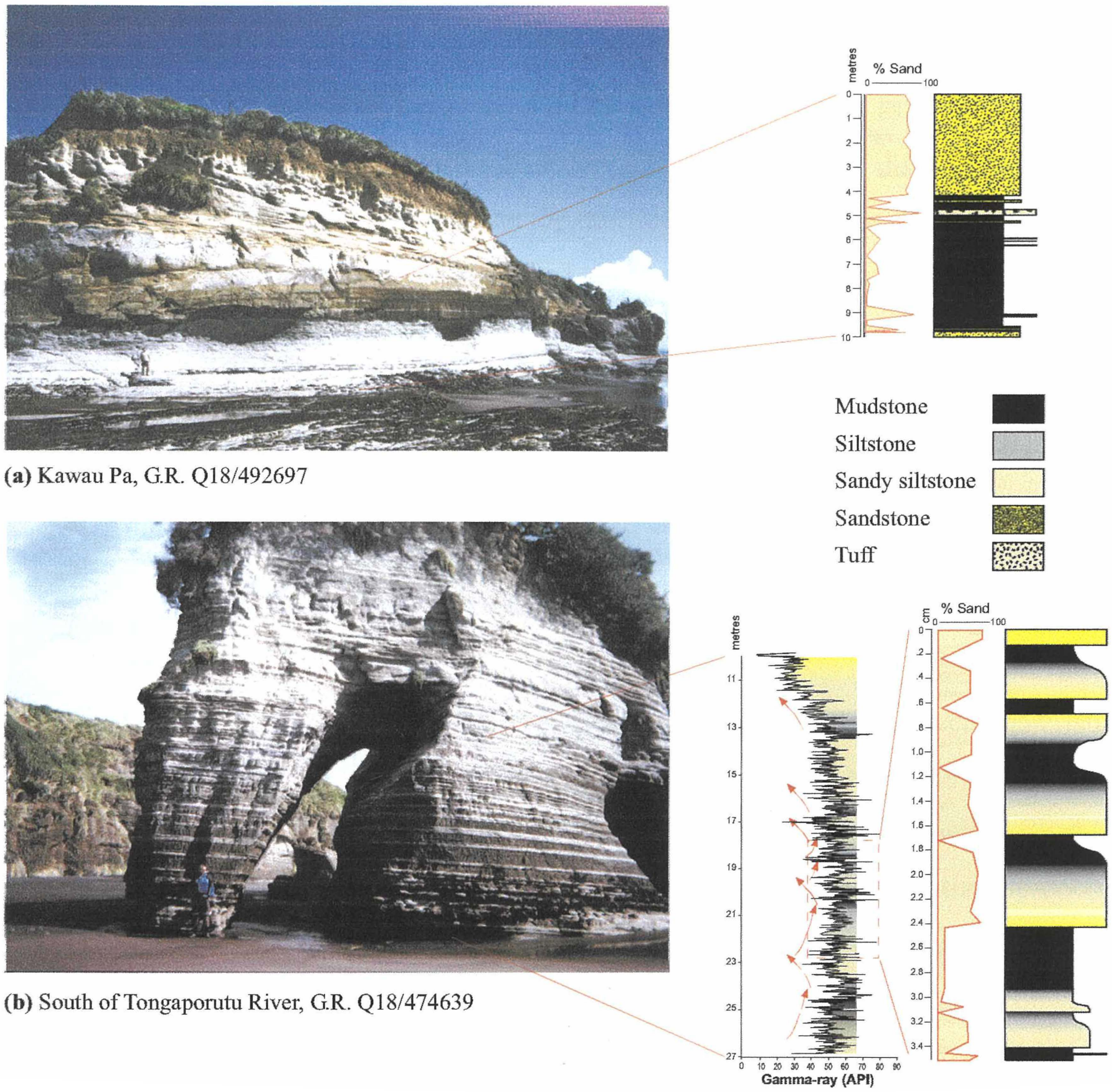


Fig. 4.7: Lower Mount Messenger Formation, displaying thick-bedded (basin floor fan lobe - top photo) and thin-bedded (fan fringe - bottom photo) facies. Note sharp basal (erosional) contact between underlying mudstone, and thick-bedded sandstone. Fining-upward character of thinner-bedded interval is observable in outcrop, textural analysis, and wireline log motif (wireline log from Jordan et al., 1994; wireline and textural positions are only approximate and representative). Basin floor fan facies of the Mangaa Formation are interpreted to have been deposited in a similar depositional environment, and are likely to display similar characteristics (bedding style, texture etc.) as the Mount Messenger Formation.

Outcrop studies of bed heterogeneity within the Mount Messenger Formation (Browne et al., 1996; Browne and Slatt, 1997; Coleman et al., 2000) established that the channel levee-overbank strata of the upper Mount Messenger Formation appear to vary vertically and horizontally more than basin-floor strata, with marked thickness variations over relatively short horizontal distances. Beds were found to be more continuous if they were greater than 35 cm thick, and over a distance of 200 m, only about 30% of sandstone beds could be correlated (Browne et al., 1996). These findings have serious implications for correlation in the study region, as the few wells in the area are very widely spaced.

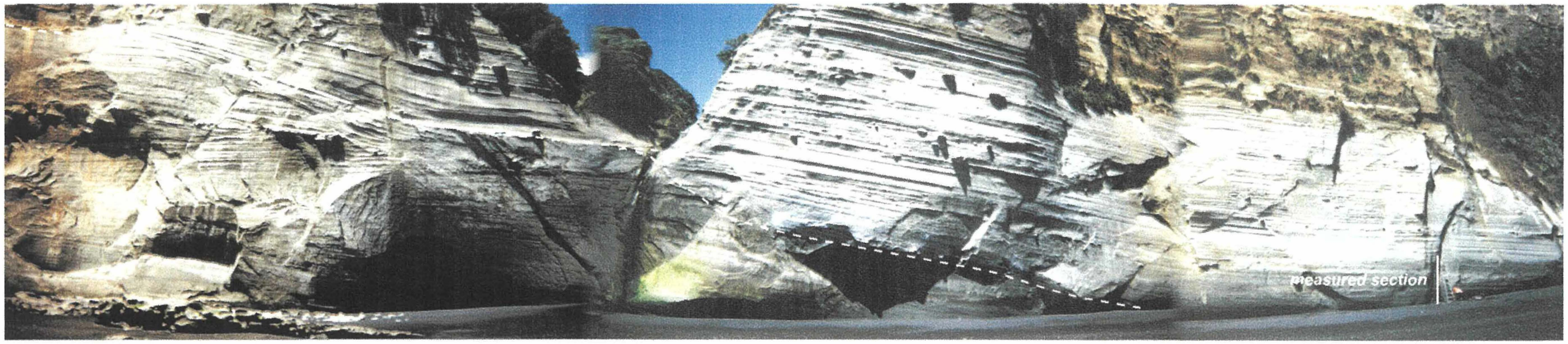
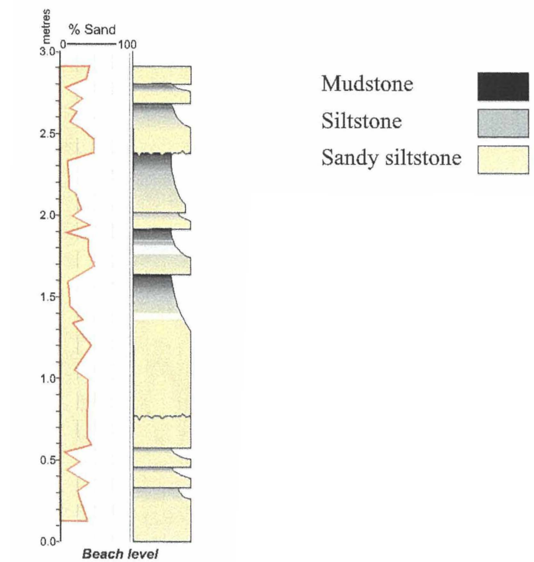


Fig. 4.8: Outcrop exposure of the upper part of the Mount Messenger Formation, Pukearuhe Beach (G.R. Q18434573). The interval is characterised by stacked, upward-fining successions, and represents turbidite deposition on channel levees or as overbank deposits on the Miocene continental slope. Dashed line defines position of channel base (after Browne and Slatt, 2002). Column (right) illustrates textural characteristics through a section of outcrop. Lithologies in general are finer-grained than displayed by lower Mount Messenger Formation strata (compare Fig. 4.7). Bedding characteristics and bed thickness of slope deposits (foresets) of the younger Giant Foresets Formation are probably similar to those displayed in outcrop. Compare also Fig. 4.13 (Giant Foresets Formation wireline character). On wireline logs, coarsening- or fining-upward units are in the order of a few metres thick. In outcrop, much more detail can be observed, with units in the order of 10's of cm's.



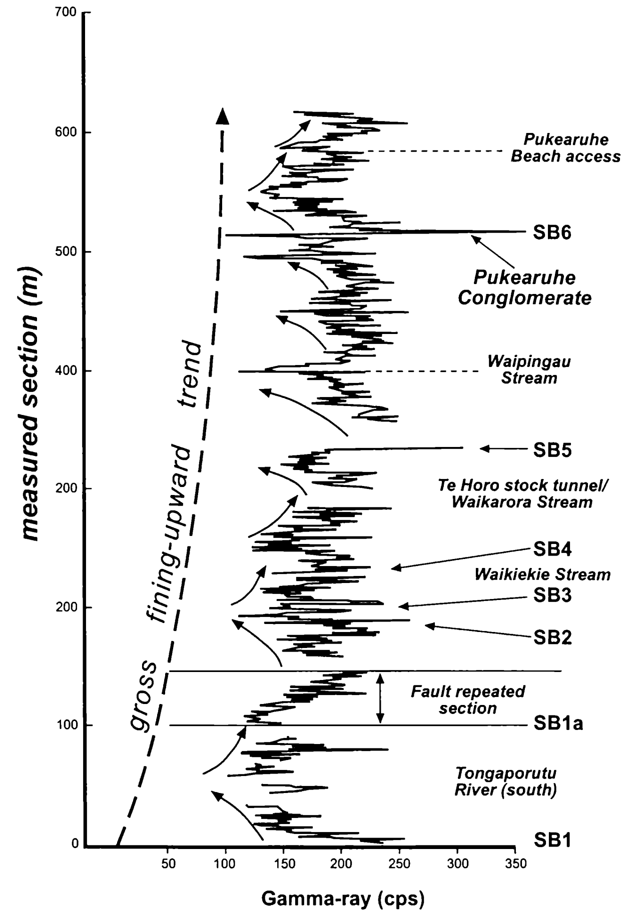


Fig. 4.9: Composite measured GR section through the upper part of the Mount Messenger Formation, northern Taranaki coastline (between Tongaporutu River and Pukearuhe Beach access road). Sequence boundaries (SB) are after King et al. (1994). From Coleman et al. (2000), modified after Diridoni (1994). Log is characterised by a succession of fining- or coarsening-upward parasequences.

4.3.4 Urenui Formation

Muddy slope sediments of the Urenui Formation are reflected in the generally quiet log signatures expressed in wireline logs. This is exemplified in Fig. 4.10 (Okoki-1), which displays a wireline signature very similar to that of the Manganui Formation. Moderate GR values and a limited degree of separation between shallow and deep resistivity logs hint at a slightly coarser texture than the Manganui Formation, while corresponding peaks on GR, resistivity, and sonic logs indicate the occurrence of shelly or conglomeritic horizons.

The Urenui Formation is interpreted by King et al. (1994) as a prograding slope complex, dominated by massive to weakly bedded mudstone, and punctuated by spectacular, isolated channel systems that are interpreted as slope feeder systems (King et al., 1993). Channels are large (up to 30 m deep and 2 km wide), and often contain a basal conglomerate, or are otherwise infilled by thick-bedded, fine to medium-grained sandstone and thin-bedded fine-grained sandstone. The muddy signature recorded both in outcrop and in subcrop strata is postulated to reflect sediment by-passing of the Late Miocene continental slope (King et al., 1993, 1994).

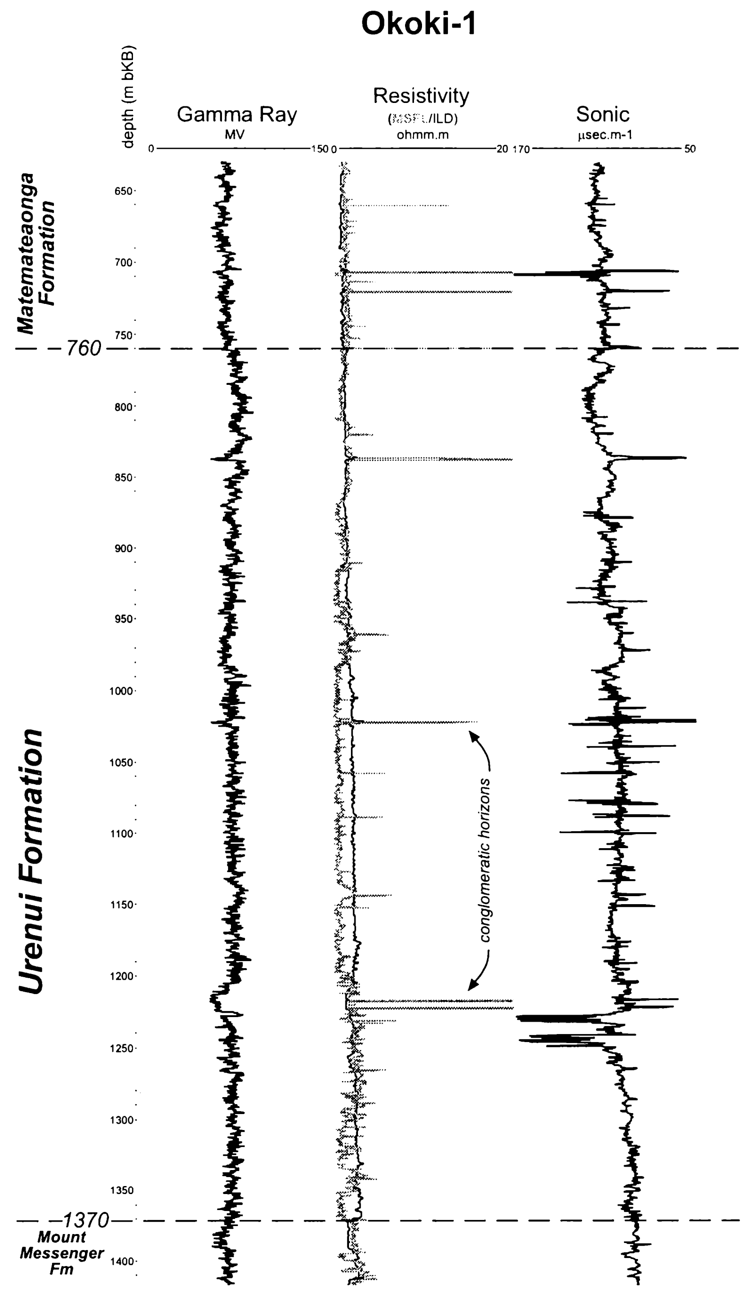


Fig. 4.10: Characteristic wireline log motifs for the Urenui Formation (example is from the nearshore well, Okoki-1). Note wireline log response to conglomeratic horizons, which line the base of large channels both in outcrop and in the subsurface.

4.3.5 Ariki Formation

The condensed Ariki Formation is a little studied but significant formation that is encountered across much of the Western Stable Platform. It is present in several wells (Ariki-1, Wainui-1, Tangaroa-1, Te Kumi-1, and also Tane-1), and has a distinctive motif that is clearly discernable on wireline logs, and easily correlatable between wells. On GR logs, the Ariki Formation may display a characteristically curvate signature, with a unique barrel-shaped profile (e.g., Tangaroa-1). Gamma-ray values shift to the left (uphole decrease) at the base of the formation, and back to the right (uphole increase) at the top of the formation (Fig. 4.11a). This signature is sometimes mirrored by a subdued SP log, and often by an increase in resistivity. Sonic and density logs differ (log values increase) from overlying and underlying formations, indicating a more condensed unit.

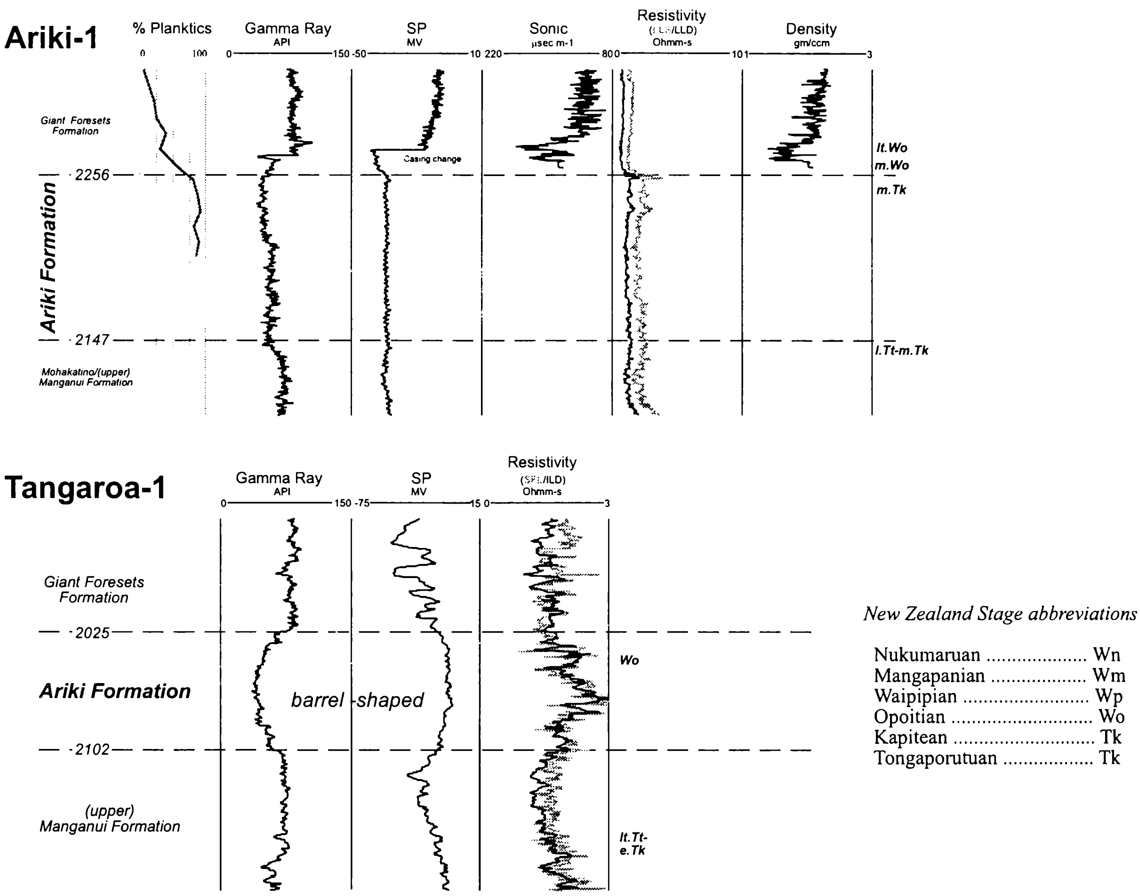
While GR values for this fine-grained formation are much lower than those observed for silts and muds, they are significantly higher than those which characterise coarser lithologies. This concurs with Villamil et al. (1998), who observed two discrete signatures on wireline logs with regard to condensed sections:

1. an increase in radioactivity on the GR curve, caused by a slow depositional rate allowing concentration of trace and rare elements relative to under-/overlying sediments, and;
2. a muted SP response indicating little or no effective or connected porosity and limited permeability.

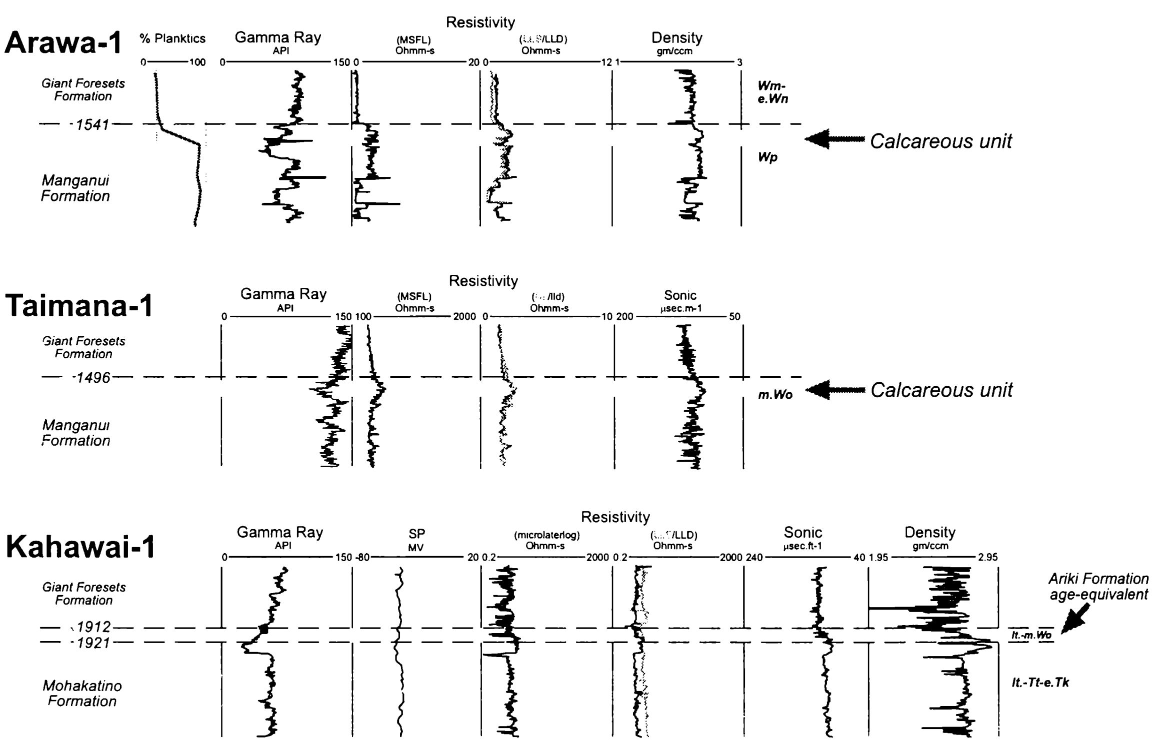
A dramatic (down-hole) increase in planktic foraminiferal content (Chapter 6) and an overall increase in calcareous fossil content relative to the Giant Foresets Formation provides an explanation for the moderate GR values displayed by the formation, as carbonate rocks record lower GR values than typically observed for fine-grained lithologies (Cant, 1992).

4.3.5(i) Time stratigraphic equivalents and other marly horizons

While the Ariki Formation is clearly defined in some wells (i.e., Ariki-1, Tangaroa-1, Te Kumi-1 and Wainui-1), similar wireline characteristics can be identified in several other wells within the study area in which the Ariki Formation has historically not been recognised. These include Arawa-1, Taimana-1, Kahawai-1, (Fig. 4.11b) Mangaa-1 and Awatea-1. At Arawa-1 and Taimana-1, a thin interval (~ 10-15 m thick) of marly siltstone, with moderate GR values, high resistivity and high sonic velocities is identified in the top part of the Manganui Formation, and is a feature used to delineate the Manganui Formation from the overlying Giant Foresets Formation. Kahawai-1 intersects the footwall of a fault block, and a very thin (less than 10 m) tuffaceous marly unit lies unconformably above Mohakatino Formation sediments.



(a) Typical wireline log reponse for the Ariki Formation



(b) Wireline log response of equivalent-age marly (condensed?) units

Fig. 4.11: **(a)** Wireline characteristics of the Ariki Formation, Ariki-1 and Tangaroa-1. Note the moderate GR, positive SP, and high resistivity, density and sonic values; **(b)** the top of the Manganui Formation often displays similar characteristics as the Ariki Formation, if somewhat thinner. The marly unit at Kahawai-1 displays a different character, probably modified by a volcaniclastic component in the unit.

At Mangaa-1, two marly-siltstone units are identified, an upper interval from ~2504-2610 m and a lower interval from 2786-2822 m. The upper interval is more silty than marly, and GR values appear to increase upward, although both SP and resistivity values indicate broadly finer-grained lithologies throughout. The lower unit, while not displaying the characteristically strong GR shift associated with the Ariki Formation, displays a weaker, yet still discernable shift to the left at the base of the interval, and back to the right at the top. These boundary shifts are also evident on the SP log. At Awatea-1 an interval between 2433 and 2540.5 m, formerly called the Urenui Formation (Murray and de Bock, 1996), and later defined as the Ariki Formation by Waghorn et al. (1996), shares the same characteristics as the upper interval described at Mangaa-1. A lower interval (~2669 m – ~2690 m) shares similar wireline characteristics, although is not described in the well report or on the composite well log. Both upper and lower intervals can be correlated between these two sites through use of seismic reflection profiles (Chapter 5), but while occurring within the approximate period as the Ariki Formation (Kapitean to Opoitian), cannot be directly correlated to the Ariki Formation at either Tangaroa-1 or Ariki-1. The thin condensed horizons noted at both Arawa-1 and Taimana-1 also cannot be correlated directly to the Ariki Formation through seismic reflection profiles.

4.3.5(ii) Significance of the Ariki Formation

Condensed and marly horizons such as the Ariki Formation and other equivalents within the northern Taranaki Basin indicate periods of sediment starvation within the study area, with lithologies dominated by pelagic sedimentation and abundant microfossil content. The importance of these horizons for correlation purposes is significant for several reasons:

1. they provide a distinct datum that is correlatable between several wells;
2. they represent a period, or periods, during which extremely low rates of clastic sedimentation prevailed, and provide evidence for significant depositional hiatus;
3. the seemingly impermeable nature of the Ariki Formation has relevance for hydrocarbon seal.

The Ariki Formation *per se* is primarily restricted to the Western Stable Platform, deposited at a time when the Northern Graben was undergoing subsidence. It represents a depositional hiatus that straddles the Miocene to Pliocene boundary, forming a significant paraconformity. The ability to correlate this event between well sections provides a constraint on the timing of events (e.g., uplift/subsidence), as well as defining which regions were experiencing active clastic deposition from those that were not. Similarly, a condensed horizon (paraconformity) is also identified at the top of the Matemateaonga Formation in the Wanganui Basin (represented by a glauconitic siltstone), and marks the informal Early-Late Opoitian intrastage boundary (Vonk et al., 2002), constraining the timing of the Early Pliocene Tangahoe pulldown (Kamp et al., 2002).

In a sequence stratigraphic context, condensed sections in marine settings often develop during transgressive and highstand systems tracts (Villamil et al., 1998). While evolution of the Ariki Formation is more complex, with both local tectonic factors (e.g., timing of subsidence of the Northern Graben), and regional factors (Early Pliocene pull-down in the Wanganui Basin) exerting significant control on the development of this formation, thinner or less significant zones of condensation may be more indicative of changes in relative sea level.

4.3.6 Mangaa Formation

The Mangaa Formation, as defined in Chapter 2, is only identified in two well sections (Awatea-1 and Mangaa-1). It immediately underlies the Giant Foresets Formation at these sites, with no apparent time break. It is an important formation, as it comprises the first major pulse of clastic sediment into the newly forming graben during the Early Pliocene. As with other thick sandstone packages in Taranaki Basin (e.g., Moki Formation, de Bock, 1994; basal parts of the Mount Messenger Formation, King et al., 1994), sandstone beds within the Mangaa Formation are exceptionally well expressed on wireline logs. The blocky to tabular profiles, bounded by abrupt upper and lower contacts, are boldly expressed on both SP and resistivity logs (Fig. 4.6), particularly at Mangaa-1. Each blocky package is separated from the next by a thin interval of finer-grained lithologies. Within each of these packages, GR, SP, and resistivity logs reveal the presence of interbedded siltstone or mudstone, indicative of thick-bedded and amalgamated coarse clastic-dominated deposits. Individual sandstone beds may be up to 10 m thick and amalgamated packages up to 40 m thick, separated by muddier lithologies in the order of 10-20 m thick. This profile mimics the log profile for the Mount Messenger Formation, and as shown in Fig. 4.6, bedding thickness are comparable. As with the Mount Messenger Formation, the log shapes are indicative of high-density turbidity current or mass-flow deposition on submarine fans.

The Mangaa Formation can be subdivided into two distinct parts on the basis of wireline character. The lower sandstone packages (Moki C2 sands of Forder and Sissons, 1992, Mangaa B sands of Murray and de Bock, 1996) display a much blockier to cylindrical shape than the upper (C1/Mangaa A) sands, which are more tabular and intercalated with thin mudstone beds (note that these sandstone beds are less developed at Awatea-1). Lower and upper sandstone intervals are separated by a thick (~100 m) siltstone at both Mangaa-1 and Awatea-1 (Fig. 4.12) and at both sites the lower part of the C1/Mangaa A sandstone displays a thinner-bedded, upward-coarsening or upward-fining SP motif, before presenting the more typical tabular motif associated with this formation. The overall log characteristics for the formation attest to two distinctive periods of basin floor fan progradation and aggradation, centered on the Mangaa-1 site; Awatea-1, as evidenced by the less well developed wireline motifs, was more distal to the fan apex.

Restriction of the Mangaa Formation based on wireline character to two wells only suggests that deposition of this formation was extremely localised, possibly fault controlled or related to a single point source (compared to a probable line source for the Mount Messenger Formation - this aspect will be explored in later chapters). Few other wells display wireline characteristics, or indeed have been lithologically described as including sandstone beds of any importance at the same stratigraphic interval as the Mangaa Formation. Te Kumi-1 and Tangaroa-1, the closest wells to Awatea-1 and Mangaa-1 (not withstanding Kahawai-1, which is located on a fault block which was undergoing uplift during this period), display a modest interval of several coarsening-upward cycles immediately overlying Ariki Formation sediments (Fig. 4.12). These may be related to Mangaa deposition on the basis of age, and are possibly distal fan-edge deposits, or may be related to some other depositional system entirely. The Late Kapitean to Opoitian sediments described in the upper part of the Manganui Formation at Arawa-1 and Taimana-1 may also be related, possibly representing a precursor to fan deposition further north. Some recent workers in fact have associated this upper sandier interval with the Mangaa Formation (G. Thrasher, pers. comm., 2002), mapping it northwards for some extent.

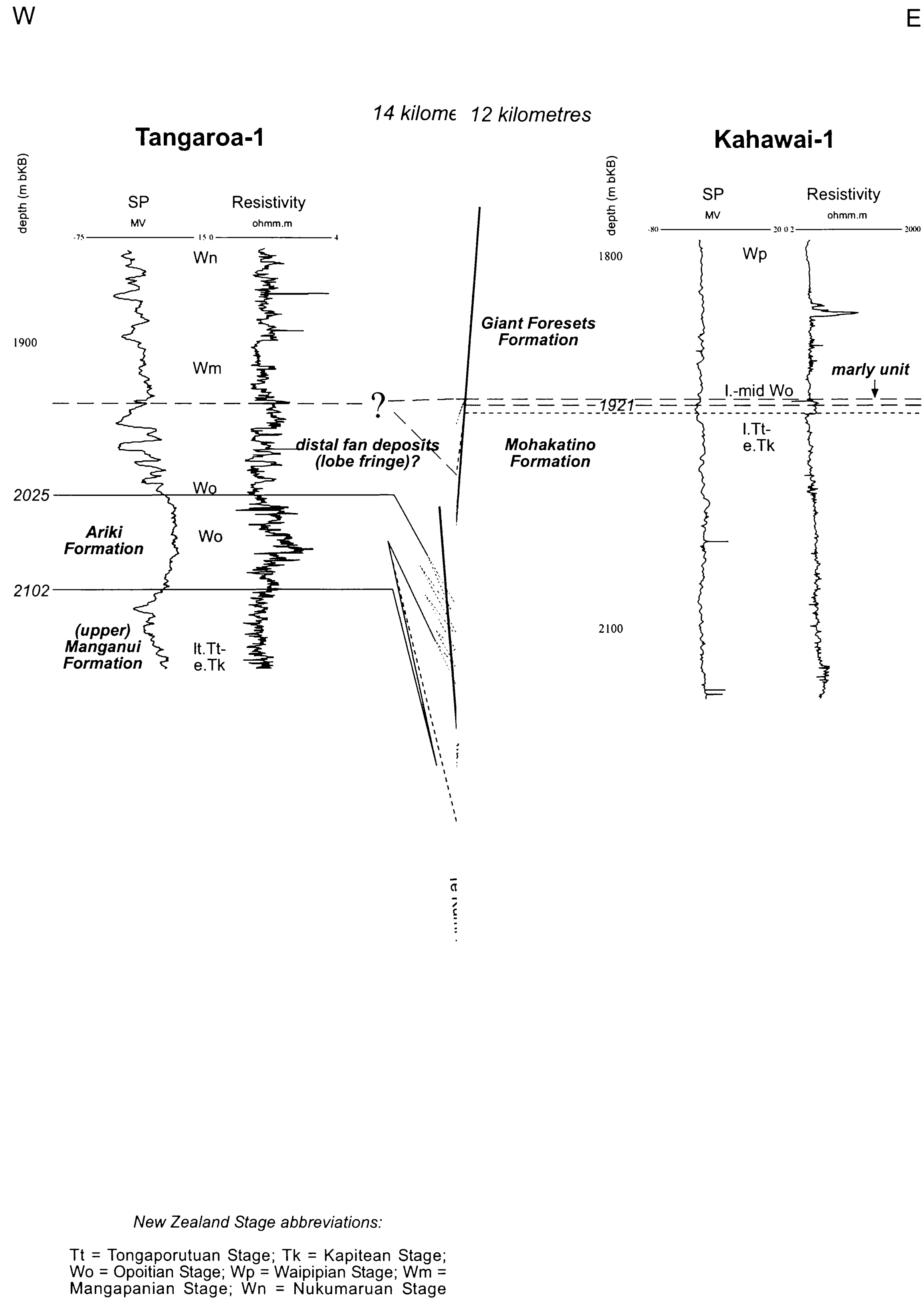


Fig. 4.12: The Mangaa Formation thins dramaticall
Northern Graben. This may be a result of lateral pinc
the fault-bound nature of the Northern Graben itse
Mangaa-1, implying that the locus of deposition was

4.3.7 Giant Foresets Formation

Although the Giant Foresets Formation is of primary significance to this study, this formation displays the least distinctive or characteristic wireline signature of any interval of interest. The predominantly muddy nature of the formation results in overall moderate to high GR values, low resistivities, with little separation between shallow and deep logs, and quiet sonic logs (Fig. 4.13). SP logs tend to be relatively featureless throughout this interval. Gross trends include an overall coarsening-upward profile, illustrated principally by a gradual transition to lower API values on GR logs.

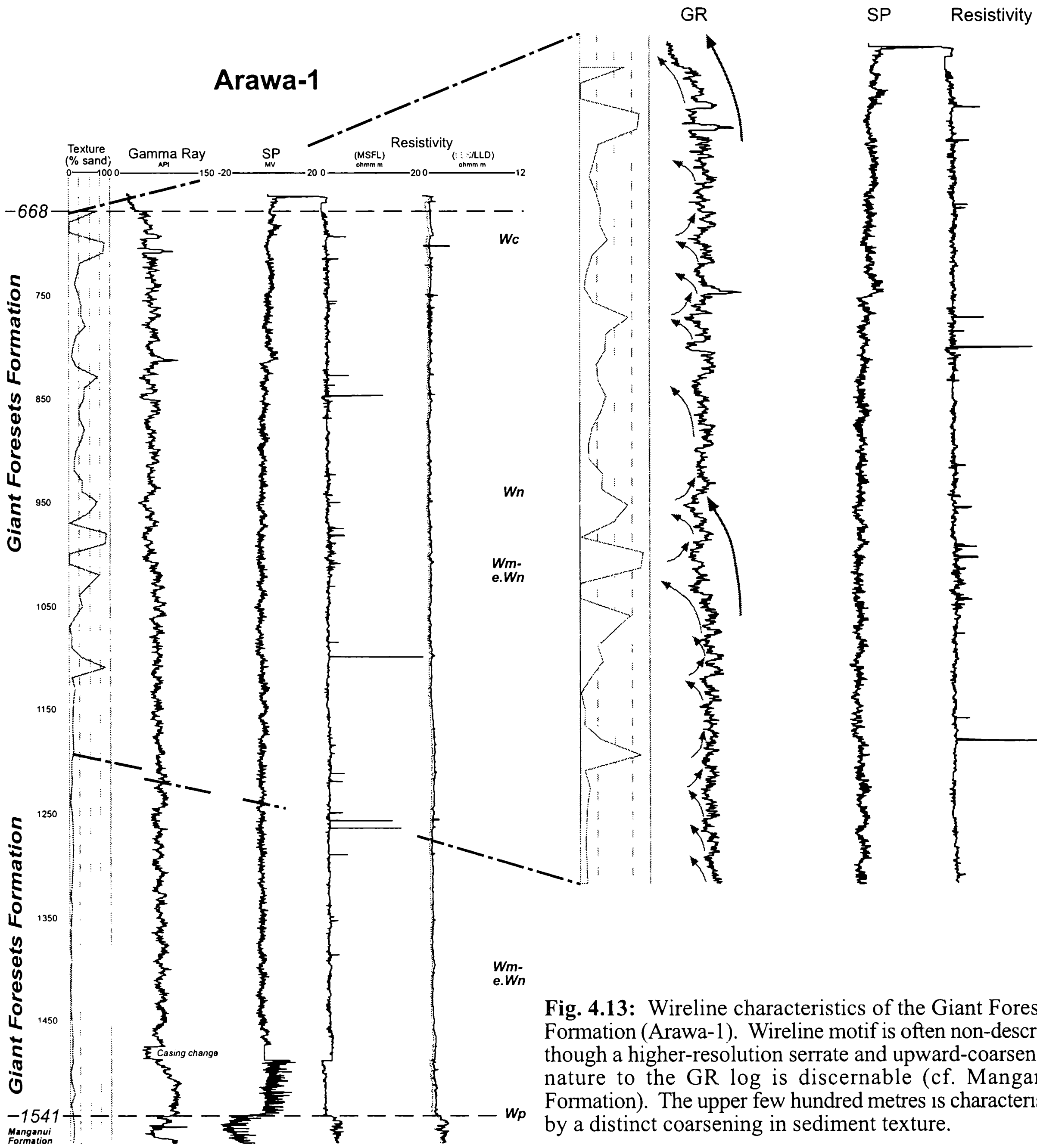


Fig. 4.13: Wireline characteristics of the Giant Foresets Formation (Arawa-1). Wireline motif is often non-descript, though a higher-resolution serrate and upward-coarsening nature to the GR log is discernable (cf. Manganui Formation). The upper few hundred metres ıs characterised by a distinct coarsening in sediment texture.

While the formation is unremarkable when compared to older underlying formations, it does however display some finer scale features, particularly in the GR log. These include small-scale fining- or coarsening-upward packages (lithologically, from claystone to siltstone to fine-grained sandstone or vice versa), with occasional discrete coarser-grained units. Sporadic thin condensed or cemented layers, possibly shell lags, are indicated by resistivity and sonic velocity spikes, which occasionally correlate with textural spikes (see, e.g., Arawa-1, 830 m bKB, Fig. 4.13). The upper 300-500 m of logged section is progressively coarser than the lower part of the formation. The increasing prevalence of serrate coarsening- or fining-upward cyclic packages uphole is supported by textural curves obtained for Arawa-1, Ariki-1, Kora-1 and Mangaa-1 (Appendix 2), and reflects an increasing terrigenous component, including lenses or pods of shell hash and other biotic debris, and occasional conglomeratic material, up through the succession. When compared with seismic reflection profiles (Chapter 5), the coarser upper part of the Giant Foresets Formation invariably corresponds to either foresets or topsets (i.e., shelf or slope facies).

4.3.8 Correlation between Taranaki and Wanganui Basins

Taranaki, Wanganui, and King Country Basins contain similar Neogene sedimentary successions, and much of their geological history is held in common (Kamp et al., 2002). Events (e.g., tectonism) in one basin have produced effects (e.g., sediment starvation/ deposition) in another. Recent work undertaken in documenting the stratigraphy, cyclicity, and depositional architecture of the Wanganui Basin succession based on wireline log motifs and outcrop study (Wanganui Research Group, University of Waikato; Griffin, 2001; Kamp et al., 2002; Kamp et al., in prep.) has accomplished much in terms of clarifying the spatial and temporal link between the Wanganui and Taranaki Basin depocentres, and also that of the recently defined King Country Basin (Kamp et al., 2002). The link between Wanganui and Taranaki Basins can be demonstrated through wireline correlation between several petroleum well sections. This link is illustrated in Enclosure 1, which displays the relationship between basins in both time and space, and also presents a schematic interpretation of the depositional paleoenvironments.

Enclosure 1 is oriented roughly south to north, beginning at Parakino-1 and proceeding across the central Taranaki Peninsula, before passing off the northern coast via Onaero-1 and Okoki-1 and into offshore northern Taranaki Basin. While the effects of Late Pliocene-Pleistocene tectonism are evident, this enclosure illustrates the genetic unity between Taranaki and Wanganui Basins, and demonstrates development of progressive Neogene sedimentary wedges. As discussed in Chapter 2, four major 2^{nd}-order transgressive-regressive megacycles of tectonic origin are recognised within the Wanganui Basin. The younger two of these, the middle to Late Miocene Whangamomona Group/megacycle, and the Pliocene-Pleistocene Rangitikei Supergroup/megacycle (Kamp et al., 2002) are of interest to this study, as the evolution and distribution of each megacycle have affected and contributed to the Late Miocene to Recent depositional architecture within northern Taranaki Basin.

The Whangamomona Group accumulated within a progradational shelf to basinal continental margin. During the middle Miocene there was extensive flooding of King Country Basin with onlap on to basement in Wanganui Basin, as well as the Patea-Tongaporutu High (Kamp et al., 2002) (Enclosure1, panels **(a)** and **(b)**. In Wanganui and King Country Basins, the Whangamomona Group/megasequence is represented by the Otunui to Matemateaonga Formations, while equivalent units in Taranaki Basin include the Manganui to Matemateaonga Formations (Fig. 2.7 and legend on Enclosure 1). It represents the northward progradation of an older progradational wedge, mainly confined to Wanganui and King Country Basins, with limited extent as a shelf-slope system along the eastern margin of Taranaki Basin. The Otunui Formation is considered to represent the thin transgressive part of this sequence, while the Mount Messenger to Matemateaonga Formations are associated with a much thicker regressive component.

Bottomset and foreset components of this progradational sequence are observed cropping out along the northern Taranaki coastline (Mount Messenger and Urenui Formations; Wai-iti Group of King and Thrasher, 1996). Wireline logs from coastal well sections show that the Mount Messenger Formation does not extend much further east than Turi-1, while the Urenui Formation cannot be traced further north than Okoki-1. Age-equivalent units of this latter formation have been subsequently eroded along the eastern margin of Taranaki Basin (e.g., Turi-1 (Appendix 2), Pluto-1). Throughout much of northern Taranaki Basin, age-equivalent sediments are incorporated into the ubiquitous fine-grained Manganui Formation. Paleoenvironmental information from composite well reports (various authors), together with the largely muddy lithology of this formation, highlight the dominantly bathyal water depths prevalent in northern Taranaki Basin during deposition of the Whangamomona Group, and are testament to the limited extent of the progradational wedge at that time. While Mohakatino

Formation sediments within northern Taranaki Basin often display a blocky wireline motif, the preferred depositional mechanism is one of mass movement (debris flow and turbidite deposition) down the flanks of volcanic massifs, rather than the basin floor fan deposits of shelf-derived siliciclastic sediment. Thus the Mohakatino Formation is technically not part of the Whangamomona progradational sequence but rather, interfingers with sediment of this group near the western limit of its extent (Enclosure 1).

The Kiore Formation and Matemateaonga Formation comprise the shelf (topsets) to uppermost slope (foresets) components of this progradational succession. As shown on panels **(a)** and **(b)** (Enclosure 1) these units do not progress far into western and northern parts of Taranaki Basin and the Kiore Formation is not recognised further west than Tuhua-1. This is partly because of its only recently erected formational status (see Vonk et al., 2002), with wells north of Tuhua-1 not yet being re-examined (note that panels **(a)** and **(b)** indicate that Matemateaonga Formation sediments identified at Onaero-1 and Okoki-1 may be better placed in the Kiore Formation), and partly because of the Late Miocene to Pleistocene erosional unconformity between peninsula Taranaki Basin and offshore northern Taranaki Basin (Manganui Platform) as a result of uplift and tilting of the eastern Taranaki Basin margin.

The accumulation of the Ariki Formation, prevalent over much of the Western Stable Platform, and various age-equivalent units, can be attributed to a number of factors or causes. The Late Tongaporutuan-mid Kapitean age for the basal part of this formation at some sites (e.g., Wainui-1, Ariki-1, Tane-1) indicates that sediment starvation in northern and western Taranaki Basin was initially associated with the restricted extent of terrigenous sedimentation associated with the Whangamomona Group. However, the uppermost part of the Ariki Formation at all sites at which it is identified is Late Opoitian to Waipipian in age, and the formation itself is invariably associated with a paraconformity. The latter parts of paraconformity (condensation) development are attributed to extensive marine flooding in Wanganui Basin, reflecting pull-down of the main Wanganui Basin depocentre during the earliest Pliocene, and concomitant sediment entrapment (Tangahoe Mudstone) within this depocentre.

The Tangahoe Mudstone is a thick (> 1000 m) mudstone succession deposited during the Early to Mid Pliocene, and marks the onset of a new progradational phase (Rangitikei Supergroup/ megacycle). The relatively rapid deposition of the Tangahoe Mudstone in Wanganui Basin contrasted dramatically with the slow sedimentation over the greater part of offshore northern

Taranaki Basin (Ariki Formation/paraconformity development). Only in the central part of the Northern Graben depositional sink, formed through extension related to the changing plate boundary configuration, did terrigenous sedimentation actively take place, with deposition of the sandy Mangaa Formation. Provenance studies (e.g., Forder and Sissons, 1992) indicate that this formation was sourced both from northern and southern landmasses, although it is highly probable that it was sourced from erosion of the Whangamomona Group in King Country Basin as a result of Pliocene doming and erosion. The Whangamomona Group sediments were themselves derived form South Island basement sources (Kamp et al., 2002).

The Plio-Pleistocene progradation that infilled Wanganui Basin advanced on two fronts, one northward into King Country Basin and the other west into the Central and Northern Grabens of Taranaki Basin (Kamp et al., 2002). Rangitikei Supergroup units thus form the lateral equivalents to the Giant Foresets Formation. Continual progradation of the Giant Foresets Formation through the latest Pliocene (with deposition of muddy bottomsets), but particularly during the Mangapanian to Recent (foresets and topsets), has resulted in outbuilding of the modern day continental shelf to its present day configuration within the study area.

Importantly, it is now understood that tectonic events in Wanganui and King Country Basins through the Middle Miocene to Pleistocene have impacted significantly on the depositional architecture and evolution of sedimentary units within northern Taranaki Basin.

4.4 Wireline facies motifs and environmental interpretation

The differentiation of sedimentary facies and their interpretation is a well-established approach to determination of depositional paleoenvironments (Walker, 1992). Facies represent the sum total of features that reflect specific environmental conditions under which a given rock was formed or deposited, be they lithologic, sedimentalogical, or faunal. This approach can also be applied to wireline logs to interpret environments of deposition. In addition, by focusing on finer-scale characteristics, and integrating these with lithological and textural information, one can interpret the depositional mechanism under which a specific facies was formed.

Gamma-ray and SP logs, despite their limitations (Appendix 4a), are often the most useful logs for interpreting facies and associated depositional environments. Typical GR and/or SP motifs used in the interpretation of facies and depositional environments are illustrated in Fig. 4.14 (see also Appendix 4a). These motifs include bell, cylindrical or blocky, and funnel shapes (Rider, 1986; Cant, 1992; Zimmerle, 1995). Other less common motifs include symmetrical or egg-shaped, and irregular or linear shapes (Zimmerle, 1995; Cant, 1992). Often, logs are paired in order to distinguish and interpret log shapes, such as GR-sonic, and SP-resistivity. Both GR and SP logs vary because of clay content (GR because of the radioactivity of clay and SP because more clay equates to less permeability) rather than because of changes in grain size (Rider, 1990). However, there is a definite relationship between clay content and grain size, and therefore between log shape and facies. Bell shapes, for example, often indicate increasing clay content (increasing shaliness, and inferred decrease in grain size) typical of channel sandstone or transgressive marine sand, while funnel shapes suggest decreasing clay content (decreasing shaliness), and a concurrent increase in grain size, indicative of barrier bar, delta front deposits, or lobe fringe/mid fan channel deposits (Rider, 1990; Cant, 1992).

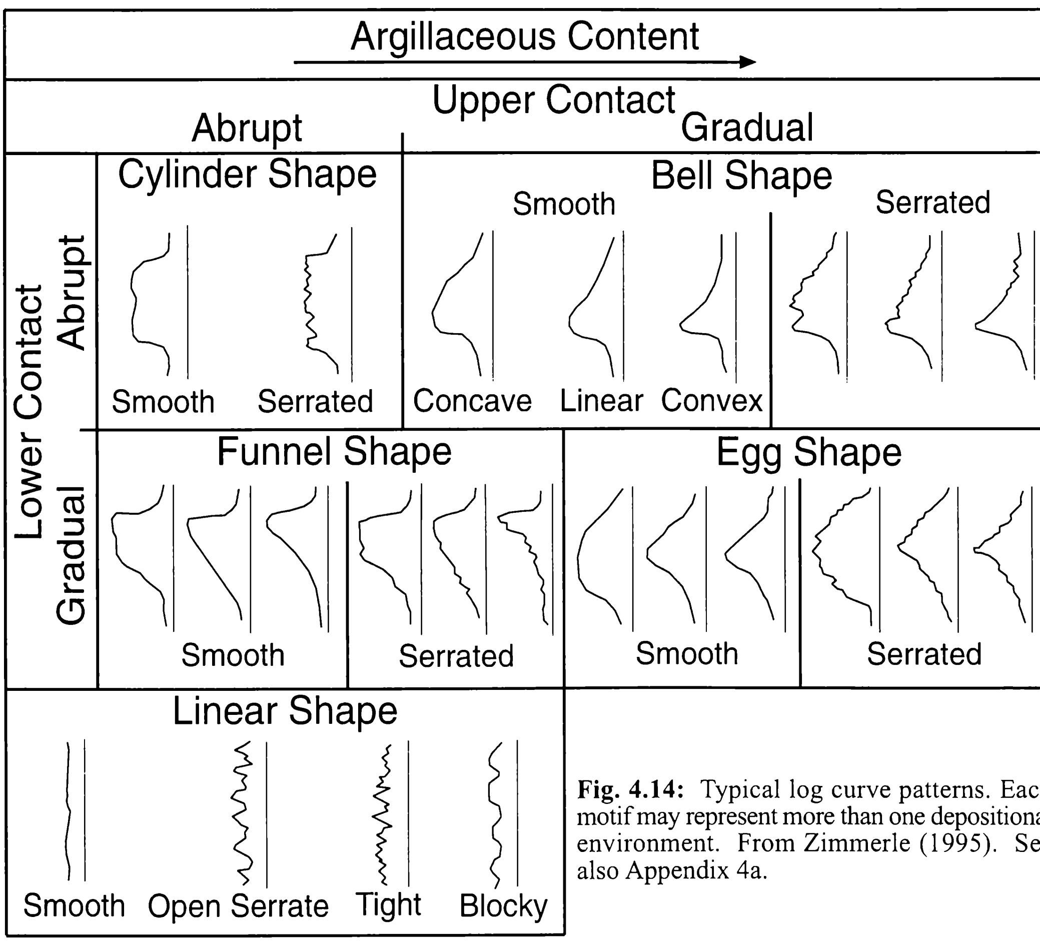


Fig. 4.14: Typical log curve patterns. Each motif may represent more than one depositional environment. From Zimmerle (1995). See also Appendix 4a.

Caution must be exercised in interpreting log shapes however, as a single log shape may be indicative of more than one environment of deposition. Therefore when any interpretation is made, it should be put into context of the known geology and paleoenvironmental conditions (Rider, 1986; Cant, 1992).

This section outlines facies interpreted on the basis of characteristic wireline motifs (integrated with lithological data), and probable depositional environments and mechanisms of emplacement. A summary of each of the wireline facies described below is presented in Table 4.2.

4.4.1 Hemipelagic (basin floor) facies

Typical wireline motif

A homogenous linear wireline motif, displaying no other specific characteristics, is typical of fine-grained basinal marine suspension sedimentation (e.g., Leeder et al., 1990), and is characteristic of much of the Manganui Formation, and the lower part of the Giant Foresets Formation. This facies is also intercalated with the sandstone beds of the Mount Messenger Formation and Mangaa Formation. GR values tend to be moderate to high (Fig. 4.15), shifting higher with increasing (down-hole) depth (compaction results in an increase in the amount of radioactive material per unit volume; Doveton, 1994). Sonic velocities also shift to the right (increase downhole) due to increasing compaction. Occasional spikes indicate more condensed horizons. The GR log is tightly serrate to almost smooth.

Lithology

Homogenous mudstone with subordinate siltstone and sandstone dominate this facies. Occasional condensed horizons, described variously in well reports as shell lags, thin limestone, or cemented layers, are also present.

Environment of deposition

The featureless nature of this wireline facies means that it is of little use in determining depositional environments (King and Thrasher, 1996). However, paleontological studies (e.g., Hayward, 1984, 1985b; Crundwell et al., 1992) conclusively indicate that the Manganui Formation was deposited at bathyal depths (greater than 200 m), while the lower part of the Giant Foresets Formation was deposited at upper bathyal to mid shelf depths. On seismic reflection profiles (Chapter 5, see also Appendix 6) this facies often correlates to horizontal/sub horizontal seismic packages indicative of basinal plain sediments. This implies deposition in a relatively stable environment, in low energy conditions, primarily as the result of suspension sedimentation.

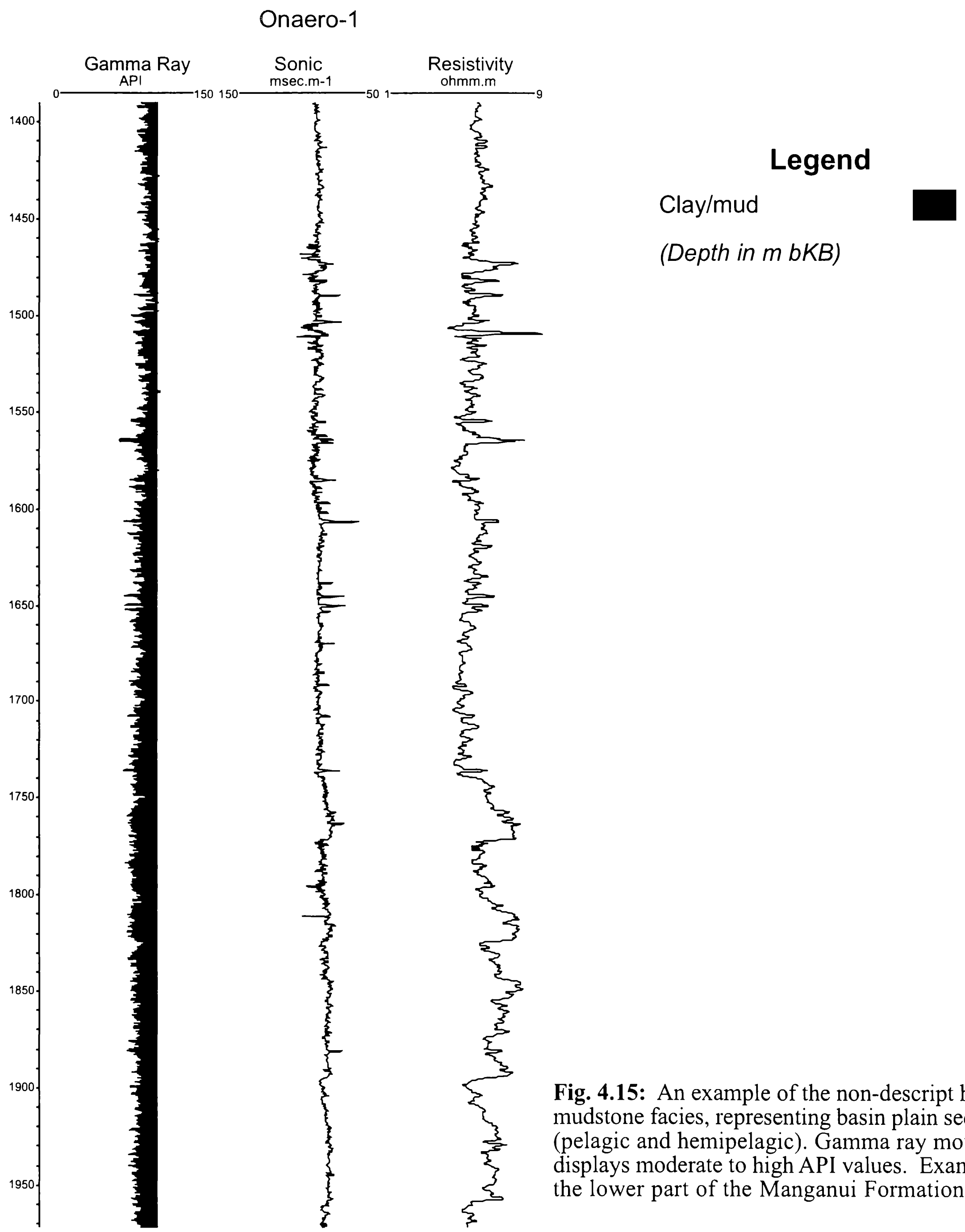


Fig. 4.15: An example of the non-descript hemipelagic mudstone facies, representing basin plain sedimentation (pelagic and hemipelagic). Gamma ray motif typically displays moderate to high API values. Example is from the lower part of the Manganui Formation, Onaero-1.

4.4.2 Basin floor fan facies

Typical wireline motif

This facies is subdivided into two subfacies. Subfacies 1 (Bff1) has wireline characteristics that include low GR values, strongly negative SP values, high resistivity, low density, and generally lower sonic velocities than muddier (dirtier) lithologies, resulting in an obvious blocky or tabular motif, with sharp and abrupt upper and lower contacts (Fig. 4.16). Each tabular package, which may obtain 50 m in thickness, is separated from the next by more radioactive lithologies (hemipelagic facies). Internally, the serrated nature of GR and SP logs within each of these packages indicates the presence of thin mudstone interbeds or amalgamation surfaces, and implies that each main sandstone body is comprised of stacked individual sandstone beds. Collectively, a succession of stacked thick-bedded or amalgamated sandstone bodies, separated by fine-grained claystone or siltstone, may attain thicknesses in the order of hundreds of metres. These tabular successions display an overall fining-upward character, with beds becoming thinner and finer-grained uphole. This facies is typical of the Mangaa Formation, and the basal part of the Mount Messenger Formation.

Discrete individual beds, a few metres to maybe 10 m in thickness, also occur sporadically throughout the succession studied, interspersed with hemipelagic facies. As with the above, these individual units are characterised by abrupt upper and lower contacts, and low to moderate GR values.

Subfacies 2(Bff2) - Often, the lower part of the Mohakatino Formation (and at some sites, e.g., Ariki-1, all of the formation) is also characterised or distinguished from overlying (thinner-bedded) volcaniclastic lithologies by a blockier log motif (see section 4.3.3, Fig. 4.3a). However, unlike the above, GR logs are more serrate, SP logs are suppressed (possibly as a result of weathered clays infilling pore spaces), and resistivity logs are much more erratic.

Lithology

Negative SP and low GR values associated with the blocky wireline motif (Bff1) are indicative of 'clean' sand-dominated lithologies, consistent with lithological comments from well reports and also from analogous basin floor fan outcrop studies (Mount Messenger Formation). These sandstone beds are interbedded with subordinate finer-grained lithologies. On seismic reflection profiles, the intervening argillaceous lithologies often correlate well with bold reflector horizons. Units ascribed to Bff2 differ slightly in that terrigenous clastic material is overprinted by voluminous amounts of volcaniclastic material, often coarsely lithic in nature (Bergman et al., 1990).

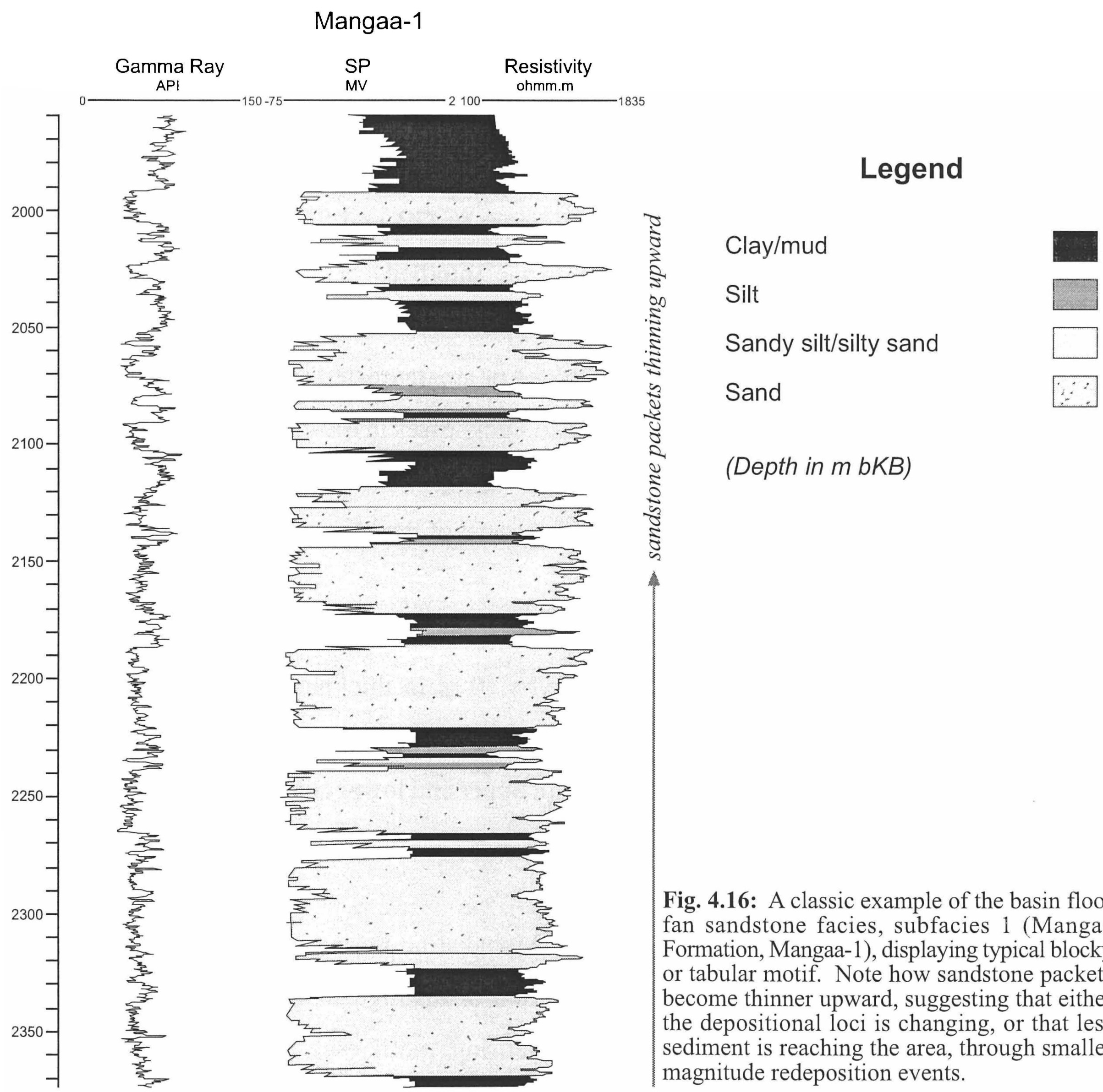


Fig. 4.16: A classic example of the basin floor fan sandstone facies, subfacies 1 (Mangaa Formation, Mangaa-1), displaying typical blocky or tabular motif. Note how sandstone packets become thinner upward, suggesting that either the depositional loci is changing, or that less sediment is reaching the area, through smaller magnitude redeposition events.

Environment of deposition

Blocky or tabular wireline motifs, and the overall broadly cylindrical shape of packages of amalgamated beds, are often indicative of high-density turbidity current or mass flow deposition on submarine fans, close to or on, the main sediment supply route and depositional axis (Walker, 1984; King et al., 1994). Corresponding mechanisms have been interpreted for the coastal (lower) Mount Messenger Formation (King and Browne, 2000). Similarly, individual units are probably the result of solitary fluidised mass flow events, such as sand-rich mass emplaced debris flows (e.g., Shanmugam et al., 1995).

The more volcaniclastic Bff2 facies are interpreted to have been the result of either mass flows (possibly triggered by successive volcanic eruptions) deposited down the flanks of volcanic edifices, or due to volcaniclastic overprint of a prograding fan system. It is difficult to determine which scenario is more plausible, although proximity to nearby volcanic massifs, and distance from the contemporary continental wedge suggests that slope failure down the flanks of these massifs is more credible. As discussed in section 4.3.3, the blocky nature becomes less obvious with increasing distance from the volcanic centre.

4.4.3 Slope facies

Typical wireline motif

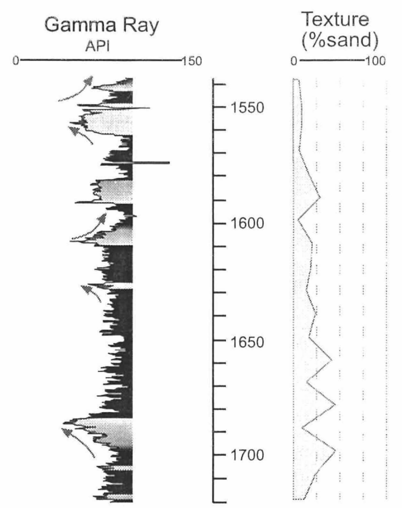
Sedimentary units typical of deposition on continental slopes commonly display coarsening-upward (funnel-shaped) or fining-upward (bell-shaped) wireline motifs, and often occur as a succession of repeated, smaller scale, cyclical packages. This motif is most often representative of turbiditic deposition. On GR and/or SP/resistivity logs, individual packages are commonly distinctly serrate, often with an abrupt upper or lower contact. Frequently, as a succession of stacked cyclical packages, they display an overall coarsening- or fining-upward trend. Four subfacies (incorporating dominant lithologies and typical wireline motifs) can be identified within this group:

Subfacies 1 (Sf1): stacked cyclical packages (each 10-20 m thick) that display moderate to low GR values, low resistivities, and low sonic velocities. Textural analysis and lithological comments indicate relatively coarse lithologies (Fig. 4.17a);

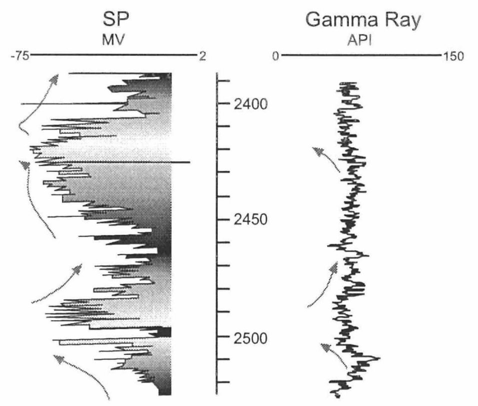
Subfacies 2 (Sf2): stacked cyclical packages that display high GR values. Dominated by muddy sediments (Fig. 4.17b);

Subfacies 3 (Sf3): fining upwards packages that display moderate to high GR values. Distinguished from previous subfacies by prevalence of volcaniclastic detritus and absence of coarsening-upward packages (Fig. 4.17c).

Subfacies 4 (Sf4) – This subfacies is subtly different to the other three subfacies but is still included in this section due to its barrel-shaped motif. It is characterised by a succession of upward-coarsening, followed by a succession of upward-fining, units that display an overall crescent or barrel-shaped profile. This motif is often evident on GR, SP, sonic and density logs (Fig. 4.18). On seismic reflection profiles, this subfacies invariably correlates to the foresetted (slope) part of the Giant Foresets Formation.

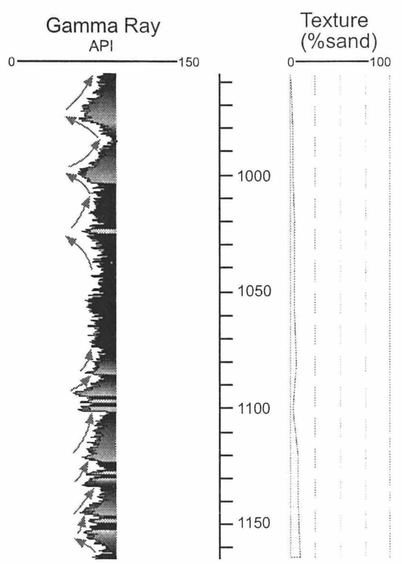


(i) (upper) Manganui Formation, Arawa-1

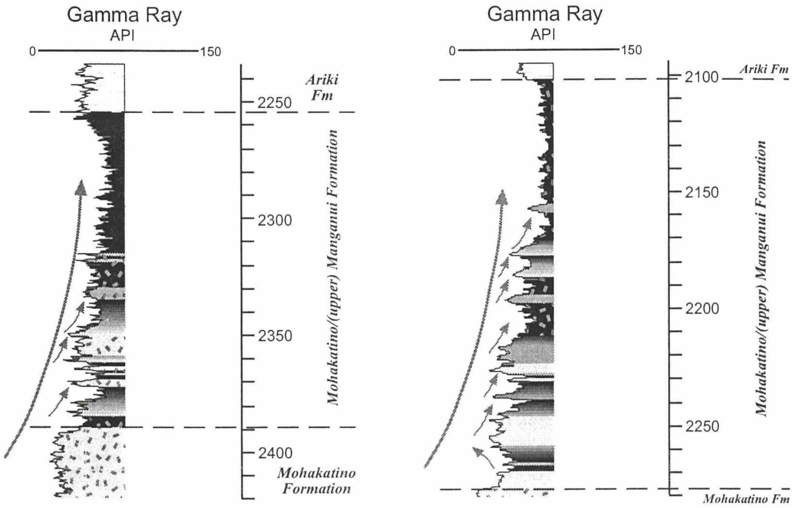


(ii) Mangaa Formation, Mangaa-1

(a) Selected examples of subfacies 1 (Sf1).



(b) Examples of subfacies 2 (Sf2), Giant Foresets Formation (Ariki-1).



(c) Examples of Subfacies 3 (Sf3), from Ariki-1 (left) and Tangaroa-1 (right).

Fig. 4.17(a-c): Wireline motifs representative of the slope facies. This facies is characterised by coarsening- and fining-upward motifs, and subfacies by log values, stacking patterns and lithology. Textural samples in (a) and (b) are averaged over 10 m intervals.

Subfacies 1 is typical of the middle to upper (foresets) part of the Giant Foresets Formation sequence, and infrequently (as at Arawa-1), the upper part of the Manganui Formation. It is also associated with the basin floor facies discussed below. The difference between Sf1 and Sf2 (common within lower-middle Giant Foresets strata) is conspicuously illustrated by the textural curves obtained for Arawa-1 and Ariki-1, (Fig. 4.17, see also Appendix 2).

Upward fining or coarsening Sf2 cycles occur wholly within fine-grained sediment, i.e., from claystone to siltstone and occasionally, to fine-grained silty sandstone/sandy siltstone (the latter two being almost impossible to distinguish between on the basis of wireline signature). Subfacies 3, almost entirely associated with the Mohakatino-Manganui Formation, displays characteristics similar to those described for the previous facies, although individual cyclical packages often display only fining-upward trends. The gross trend for this subfacies is also one of fining-upwards. The very spiky nature of all subfacies log motifs is indicative of rapidly alternating lithologies (interbedding), and suggests variable sediment supply within longer-lived sedimentary episodes.

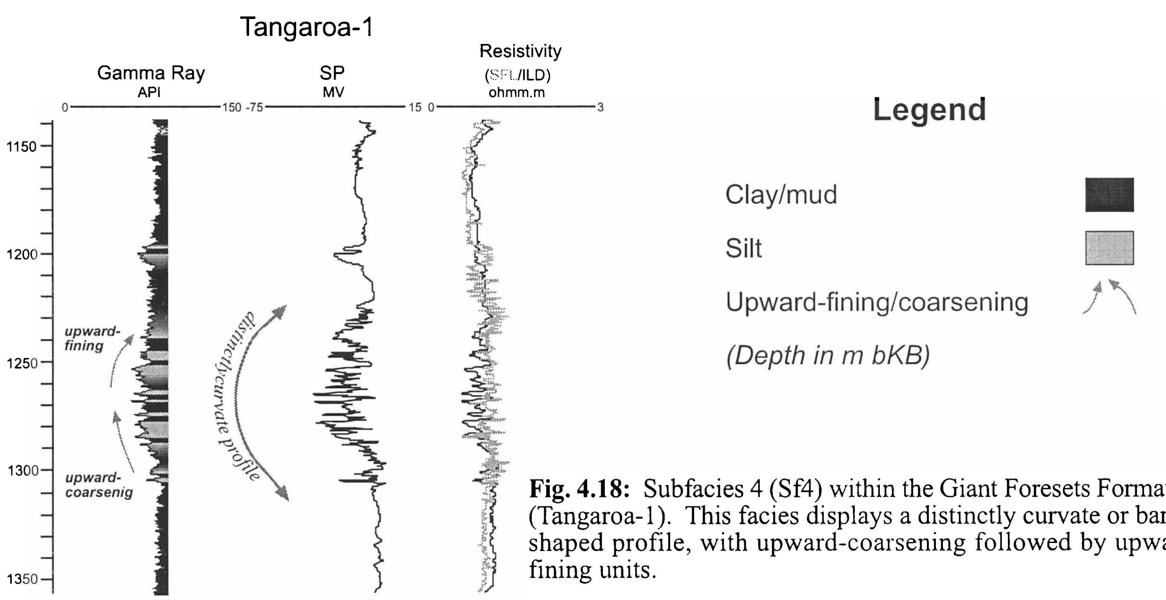


Fig. 4.18: Subfacies 4 (Sf4) within the Giant Foresets Formation (Tangaroa-1). This facies displays a distinctly curvate or barrel-shaped profile, with upward-coarsening followed by upward-fining units.

Lithology

Lithologically, these subfacies vary considerably. Subfacies 1 and 4 typically grade from clay to silt (dominant) to coarse sand. Subfacies 2 and 3 are dominated by claystone and siltstone, and occasional fine-grained sandy siltstone/silty sandstone. The lower GR values exhibited by Sf3 may be a result of the relative abundance of lithics in the sediment, or, as discussed in section 4.3.3, the result of weathering of volcanic detritus to smectitic clay minerals.

Environment of deposition

Many of the units within this facies are attributed to deposition in a channel-levee complex on the paleo-slope. Channel-levee complexes (slope fan complex of Van Wagoner et al., 1988) develop when river-derived sands are transported down incised valleys and canyons into the basin via mass flow and turbidite mechanisms. Sands are commonly confined to channels, and may be indicated by one or more massive sandstone beds with prominent sharp bases and an upward decrease in sand content (bell-shaped motif), or may have a blocky log shape somewhat resembling a basin floor fan, but with a more thinly-bedded or ragged character. Muddier sediments form overbank or levee deposits, and tend to display coarsening-upwards (funnel-shaped) or crescent-shaped profiles (Vail, 1987; Mitchum et al., 1993). Most channel-levee complexes are composite units made up of several individual leveed channel systems (Mitchum et al., 1993). This is illustrated in Fig. 4.19, which displays the variety of log patterns that can occur dependent on position within the complex.

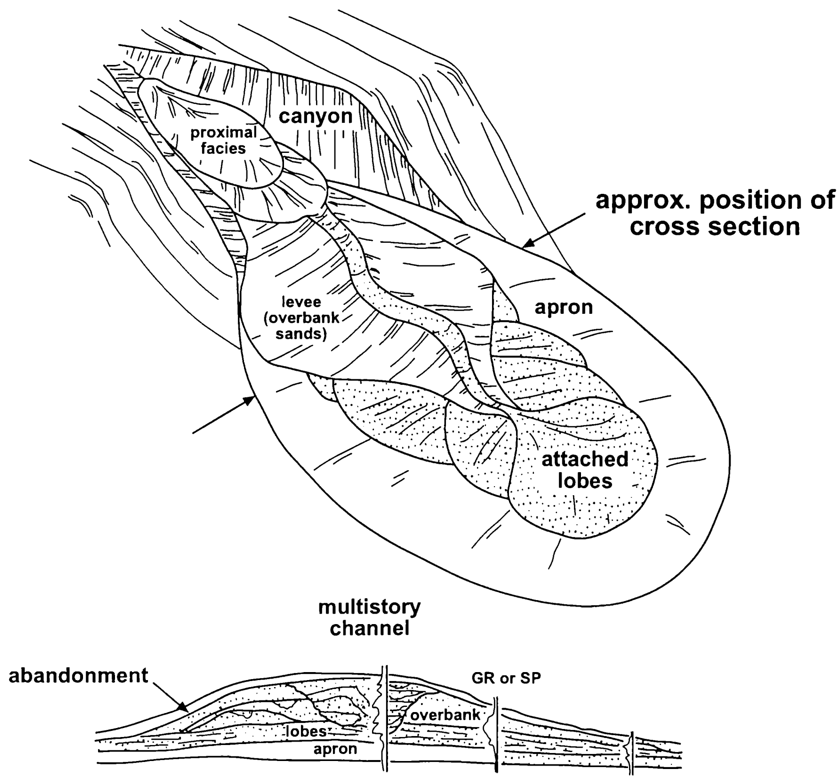


Fig. 4.19: Idealised depositional diagram of a single leveed channel system, showing development of component parts (top) and idealised well log responses through the various parts (bottom). After Mitchum et al. (1993).

In coastal outcrops, thin-bedded sandstone of the (lower) Mount Messenger Formation are interpreted as lobe fringe deposits within the basin floor fan (King et al., 1994) (Fig. 4.7). Higher in the succession, the channel-levee complex produces a GR log motif very similar to

Sf1, with fining- (dirtying) and coarsening- (cleaning) upward trends exhibited. King et al. (1994) interpret these as reflecting crevasse splay or levee overbank deposition away from the main sandstone depocentres or transport axes, and may represent the coalescence of several relatively small slope fan systems (Browne and Slatt, 2002). Low density turbidity currents account for much of the sediment deposited, though thinning- and fining-upward cycles are attributed to autocyclic depositional adjustments, such as lobe switching, compensational bedding, and channel-levee migration (Jordan et al., 1994; King et al., 1994). Similarly, most funnel-shaped Sf1 units through the Giant Foresets Formation are interpreted as levee-overbank deposits, possibly as part of a prograding channel-attached lobe system. Correlation with seismic reflection profiles (see Appendix 6) illustrates that this interpretation is reasonable, based on position in the shelf-slope-basin system. The crescent-shaped log motif associated with Sf4 is characteristic of channel-overbank units and multistory sandy intervals within the channel-levee complex (Mitchum et al., 1993) (Fig. 4.19), and as with upper Mount Messenger sediments, may in fact be the result of the interfingering of a number of smaller slope fan systems.

Funnel-shaped Sf1 packages are also associated with the basin floor fan facies of the Mangaa Formation. The coarsening-upward profiles created by Sf1, and the progressive overall coarsening-upward profile and subtle increase in the tabular nature and thickness of the stacked packets of the basin floor facies (Fig. 4.20) characterises a prograding submarine fan system (Cant, 1992). In this system, Sf1 typifies prograding fan deposition (and initial lobe switching), while the main sandstone packages (Bff1) represent major periods of fan development (mid channel fill or sheet sands), and the intervening siltstone (hemipelagic facies) represent a continuum of background sedimentation. Later fining-upward (or thinning and 'dirtying' of beds) signify progressive fan abandonment. This development-abandonment-development pattern is repeated both in the Mount Messenger and Mangaa Formations.

Subfacies 3 is interpreted to represent turbidite deposition downslope from volcanic edifices, rather than being sourced from the shelf. Nodder et al. (1990b) suggests that deposition of similar facies that crop out north of Awakino River (his Ngarupupu Formation) occurred in volcaniclastic aprons around, and away from, the perimeter of active volcanic complexes. Given the proximity of wells in the offshore northern Taranaki Basin to the Miocene volcanic massifs, it follows that this would hold true for Mohakatino Formation sediments, with sediment deposition occurring down and out from the flanks of massifs.

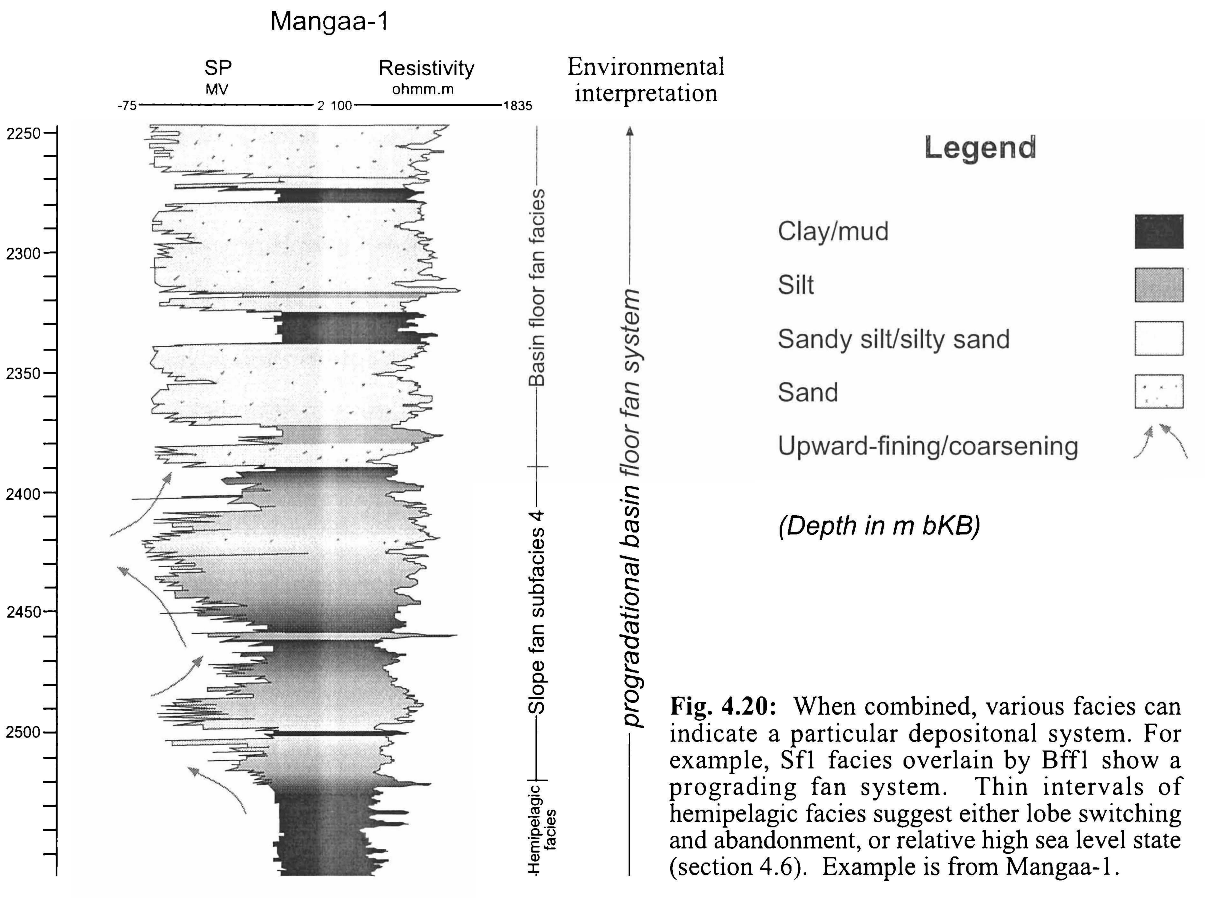


Fig. 4.20: When combined, various facies can indicate a particular depositonal system. For example, Sf1 facies overlain by Bff1 show a prograding fan system. Thin intervals of hemipelagic facies suggest either lobe switching and abandonment, or relative high sea level state (section 4.6). Example is from Mangaa-1.

Interpretation of Sf2 sediments is somewhat more ambiguous. These occur often as a series of stacked packets within the bottom sets or lower foresets part of the Giant Foresets Formation, and occasionally in the Manganui Formation, and reflect only slight changes in sediment texture (muds or clays to silts). They may represent the very distal parts of turbidites deposited on the basin floor, or may be the result of pulses of sediment related to cyclical sea-level change that are less diluted by transport distance.

4.4.4 Shelf facies

Typical wireline motif

Two subfacies can be identified within this facies division:

Subfacies 1 (Shf1): stacked cyclical packages (each 20-40 m thick) that display moderate to low GR values, low resistivity, and low sonic velocities compared to slope strata. Spontaneous potential values are also often subtly lower (more negative) than slope facies. This subfacies is very similar to Sf1 (slope facies) but can be distinguished by a greater degree of separation between the shallow and deep resistivity curves, and generally coarser lithologies (Fig. 4.21a). This facies is more often observed as coarsening-upward, rather than fining-upward, packages.

Subfacies 2 (Shf2): this facies is characterised by moderate GR values that in many cases only subtly stand out from wireline log, with a vaguely barrel- to rounded shape, and a slightly greater degree of separation between the shallow and deep resistivity curves indicating coarser lithologies (Fig. 4.21b). The often serrated nature of the GR log suggests a degree of bedding within each unit. Thicknesses range from a few metres to greater than 30 metres, but are more often less than 20 metres. At some sites (e.g., Kahawai-1, Arawa-1) these units are prevalent in the upper part of the sequence, but at most sites occur more sporadically through the upper foreset and topset part of a sequence.

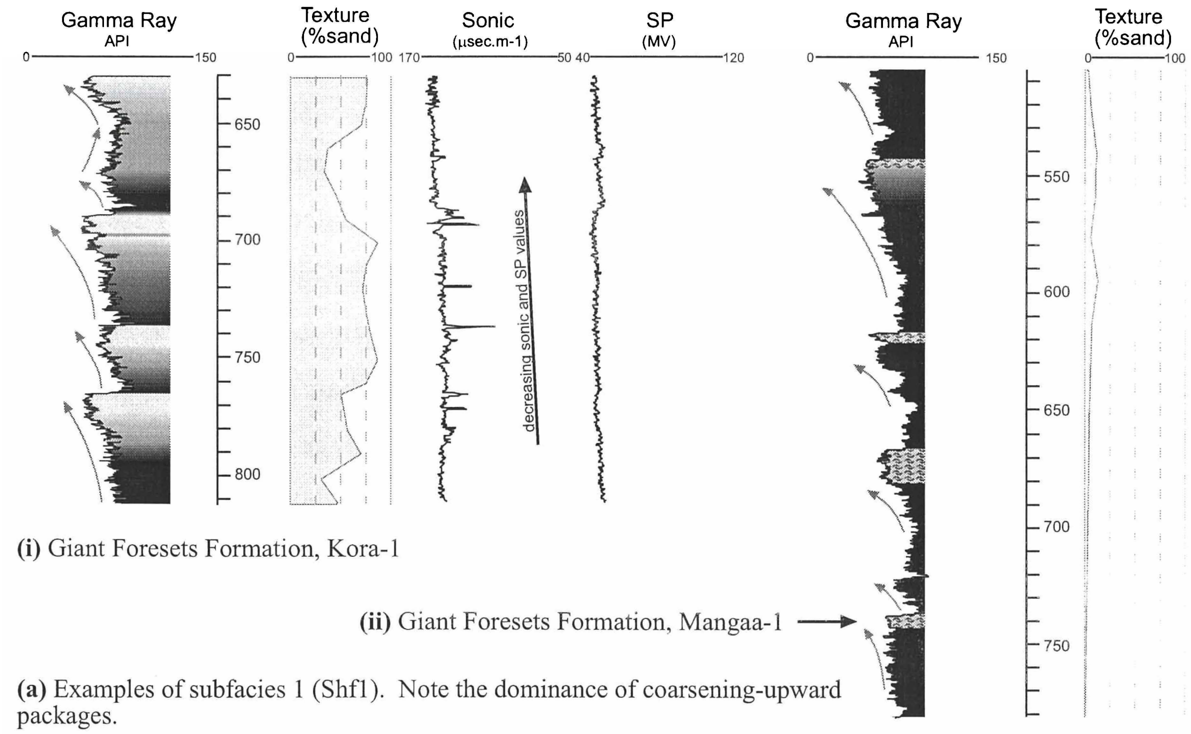
Lithology

Generally, upward-coarsening packages within Shf1 grade from mudstone to siltstone to sandstone. In a number of instances (e.g., Mangaa-1) several consecutive coarsening-upward packages are capped by a layer of shell hash, which include variable amounts of angular shell fragments (in parts shell hash is prevalent) and occasional pebble horizons (Shell BP Todd Oil Services Ltd., 1981, 1984; ARCO Petroleum NZ Inc., 1992). Some of the shell material has been identified as originating from a shallow inner to mid shelf environment (A. Hendy, pers. comm, 2002), providing evidence of transportation of sediment from the nearshore region to mid shelf and upper bathyal depths.

Subfacies 2 is lithologically variable, ranging from relatively coarse sandstone to sandy siltstone. Where textural curves have been derived (e.g., Arawa-1), these units correlate well with an increase in sand percentage.

Environment of deposition

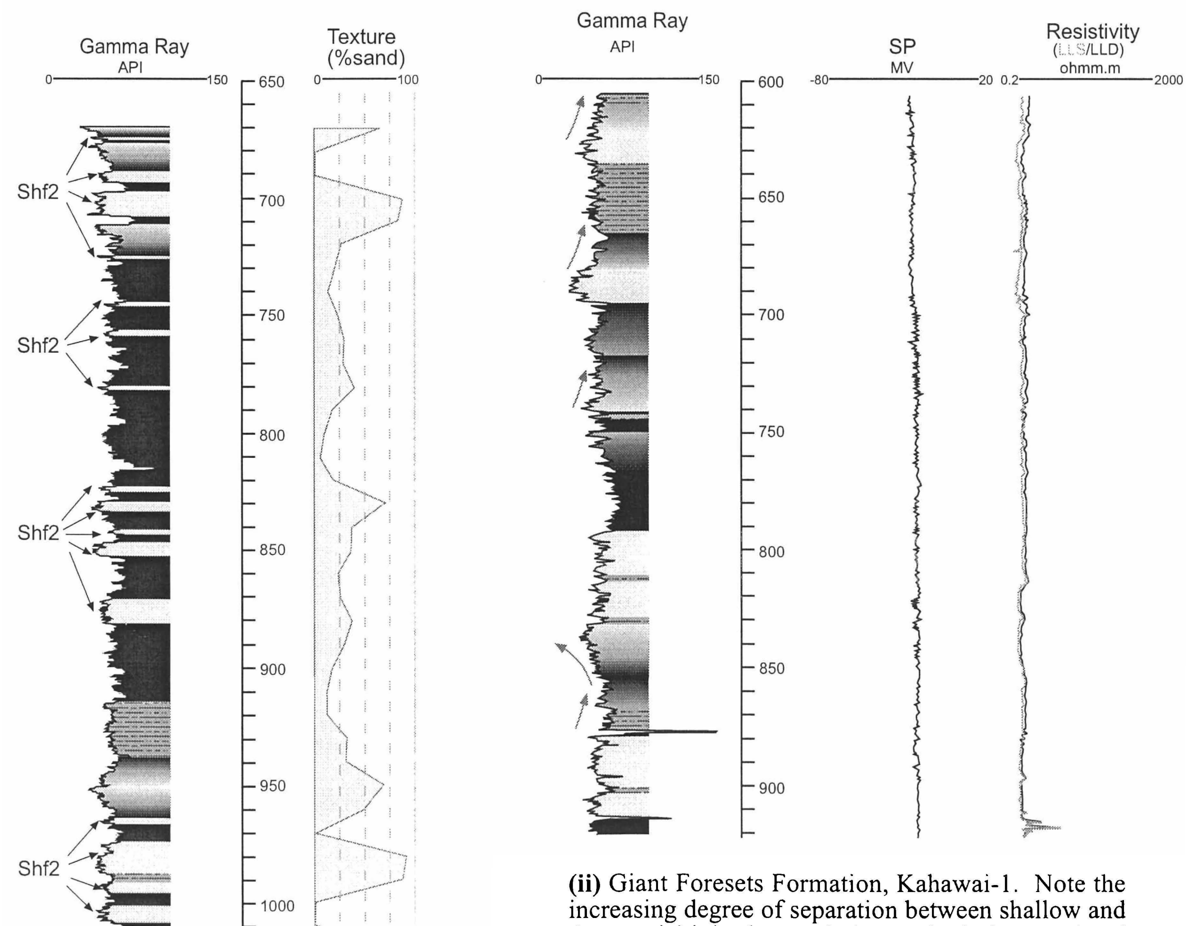
Interpretation of depositional environments can be hampered by the fact that many log patterns can be representative of a number of environments of deposition (e.g., Danielsen et al., 1997). For this facies group, interpretation of a shelfal to uppermost slope depositional environment is supported by position on seismic reflection profiles (Chapter 5) and by biostratigraphic data.



(i) Giant Foresets Formation, Kora-1

(ii) Giant Foresets Formation, Mangaa-1

(a) Examples of subfacies 1 (Shf1). Note the dominance of coarsening-upward packages.



(i) Giant Foresets Formation, Arawa-1

(ii) Giant Foresets Formation, Kahawai-1. Note the increasing degree of separation between shallow and deep resisitivity logs uphole, particularly associated with Shf1.

(b) Examples of subfacies 2 (Shf2). Textural curves, where estimated, illustrate the coarse nature of this facies.

Fig. 4.21: Wireline motifs representative of the shelf facies. This facies is generally characterised by moderate to low GR values and relatively coarse lithologies. See previous figure for key.

Funnel-shaped motifs such as described for Shf1 can be indicative of numerous environments, including barrier bars, shallow marine sheet sandstones, and submarine fan lobes (Cant, 1994). The shelfal position of Shf1 however, precludes deep submarine fan deposition or shallow marine/neritic deposition. More probable, these motifs are a result of the basinward movement of the shoreline during falling sea level, and may be considered regressive units (e.g., Naish and Kamp, 1997) deposited shoreward of the contemporary shelf break. It is difficult to ascribe a single mechanism to the deposition of Shf2. The slightly barrel or rounded shape to individual units may suggest transgressive shelf sands (Cant, 1994), though they may also be the result of channel fill, or have been deposited as the result storm surges on the continental shelf (e.g., Reynolds, 1994).

4.4.5 Slumped facies

Typical wireline motif

Wireline motifs approximate those observed for the Sf4, with moderate GR values and more negative SP values, but also have several distinguishing characteristics:

1. sonic and density values are much more erratic, suggesting more chaotic mixing of sediment;
2. while the overall log motif is distinctly crescent-shaped, it is not consistently comprised of upward-coarsening followed by upward-fining units (Fig. 4.22);
3. where slope channel facies may only be a few ten's of metres thick, slumped facies are in the order of hundreds of metres thick.

Lithology

Comments from composite well reports, where noted, indicate that these intervals are lithologically variable, and include siltstone, sandstone, and shell hash, not unlike Sf1 and Sf4.

Environment of deposition

Thick crescent or barrel-shaped GR units are suggestive of mass emplacement such as slumping of the shelf break. This facies occurs beyond the (contemporary) shelf break, on the upper to middle slope, and corresponds to the degradational foreset facies of Beggs (1990). It is interpreted to form as a result of lowering of relative sea level, during which the shoreline progrades seaward. As sea level lowers, large volumes of sediment are delivered to the shelf/

slope margin, where it accumulates and becomes overlie steep and ultimately unstable. At some point, destabilisation causes collapse of the over-steepened margin, resulting in slumping of the accumulated sediment down the slope (Dam and Sønderholm, 1994).

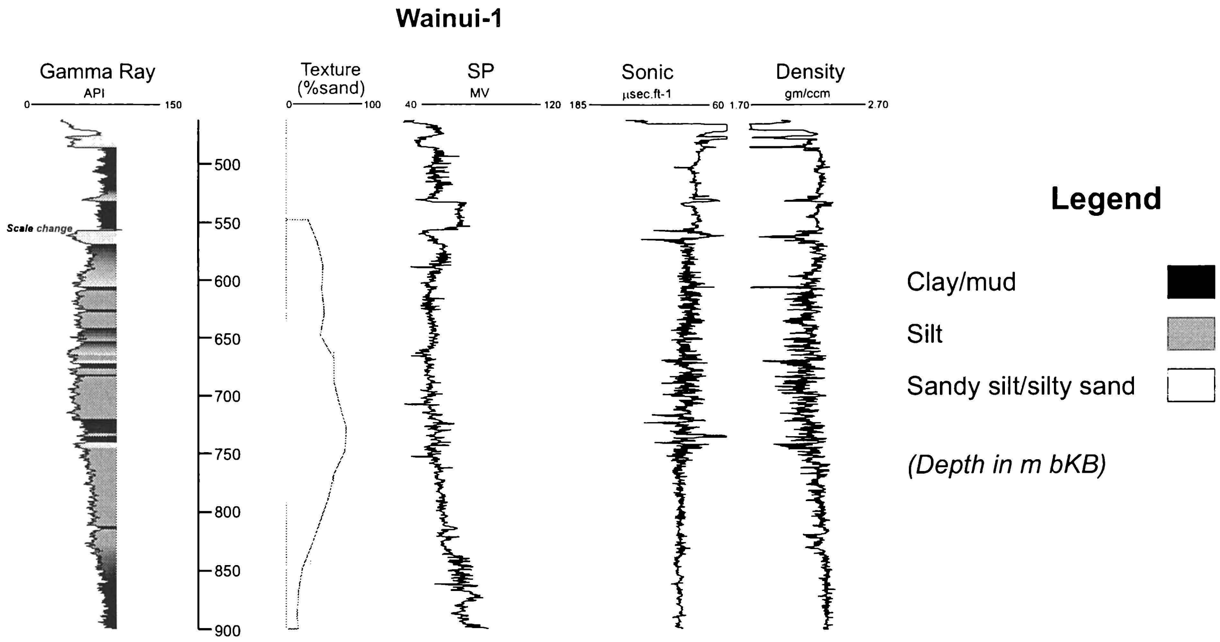


Fig. 4.22: Example of the slumped facies (Giant Foresets Formation, Wainui-1). Note the similarities between this facies and Sf4, with an arcuate profile, and moderate GR values. However, whereas the latter facies is scaled by a factor of tens of metres, the slumped facies occur in packages in the order of hundreds of metres.

4.4.6 Condensed (calcareous) facies

Typical wireline motif

The typical wireline motif of the calcareous facies has previously been defined in section 4.3.5 (Ariki Formation). Refer to Fig. 4.11(a,b) and Table 4.2 for characteristic log shapes.

Lithology

This facies is very calcareous, and in well completion reports is described as a very fine-grained siltstone to marly mudstone, containing a high abundance of microfossil (foraminifera) remains.

Environment of deposition

Condensed horizons are interpreted to form as the result of sediment starvation during periods of relatively high sea level, during which background pelagic or hemipelagic sedimentation dominates. Analysis of sedimentary sequences in the northern Gulf of Mexico led Villamil et al. (1998) to suggest that condensed horizons are relatively common in stratigraphic successions, developing to a limited extent where terrigenous sediment deposition has been abandoned (e.g., on top of sheet sands, and within levee or overbank sediments of slope fans). However, these are often not apparent on two-dimensional seismic profiles. In the Villamil et al. (1998) study, condensed sections were interpreted locally as separating slope fan facies from prograding complexes, and associated with a regional transgression and a landward shift of deltaic depocentres.

Sediment starvation is proposed as being the most significant factor in development of condensed horizons within northern Taranaki Basin. Thinner horizons such as would be associated with fan abandonment or regional transgression are difficult to identify based on wireline motifs within the study area.

4.5 Reconstruction of depositional systems based on wireline log facies motifs

In order to fully understand the relationships between facies outlined in section 4.4, one needs to put them into a wider (regional) context. This is achieved by way of a fence diagram (Enclosure 2), which illustrates both the spatial and temporal relationship between facies identified at each well site.

This figure displays several obvious features:

1. Hemipelagic sedimentation (hemipelagic facies) dominates much of the Miocene interval to the west. To the east, hemipelagic sediment interfingers with volcaniclastic dominated sediment, probably, based on age and distribution patterns, sourced from the Kora volcanic complex (~15 Ma) and later by the Mangaa and Mangaa South volcanic complexes (12.5 and 11.5 Ma respectively). The age of volcaniclastic facies intersected at both Ariki-1 and Tangaroa-1 suggest that the Te Kumi-1 complex (active about 15 Ma; Thrasher et al., 2002) and the Awatea and Mangaa complexes (12.5-10 Ma) contributed to the sequence at these sites. Thinner volcaniclastic units (Bff2 and Sf3) intersected at Arawa-1 may be the result of distal subaqueous fall, or sourced from the small Caelum volcanic centre (~13.5 Ma). The

southwesterly location of Wainui-1 suggests that volcaniclastic Bff2 facies are the result of reworking of volcaniclastic material from the northeast.

2. The intercalation of late Middle to Late Miocene quartzofeldspathic facies (Bff1, Sf1, Sf2) with the dominant volcaniclastic facies (Bff2, Sf3) at Turi-1 and Kahawai-1 suggests limited introduction of south derived coarse-grained sediment into the northern parts of the basin at this time.
3. Condensed facies are widespread over the Western Stable Platform during the latest Miocene to Early Pliocene, are older to the west and north (Wainui-1, Ariki-1), and persist to a later stage in the north (Tangaroa-1, Te Kumi-1).
4. The thick basin floor fans (Bff1) clearly evident at Awatea-1 and Mangaa-1 are extremely localised. Late Opoitian and Waipipian occurrences of Sf2 (slope facies) units observed at Tangaroa-1 and Te Kumi-1 may be interpreted as distal turbidites related to basin floor fan facies at Mangaa-1 and Awatea-1, but this is only a tentative correlation. The sandstone units are thickest and best developed at Mangaa-1, suggesting that sediment was sourced from the east, or possibly via a sediment pathway down the eastern axis of the Northern Graben. Lack of correlative units at Kahawai-1 indicates that sediment bypassed this site.
5. Other than at Awatea-1 and Mangaa-1, muddy Sf2 facies dominate Pliocene successions, with lithologies become increasingly sandy into the Pleistocene (Sf1), reflecting increasing proximity to the shelf break. The sudden increase or dominance of Sf1 and/or slumped facies heralds the arrival of the leading edge of the advancing slope.
6. Degradational foresets occur in the Late Pliocene to Pleistocene, and only in the more westerly parts of the study area. This suggests that there are intrinsic differences between earlier foreset deposition and later foreset deposition. Differences in lithology (sandier, shellier) and physical changes in mud (colour, plasticity, stickiness, as discussed in Chapter 2) may indicate a greater contribution from the more proximal North Island, rather than a dominant South Island source.
7. Sandy and shelly Shf1 and Shf2 facies reflect topset facies. These facies are indicative of shallower depths (Chapter 6) and greater connectivity between shoreface/shallow neritic environments and mid to outer shelf environments.

Correlation of individual beds, or even successive facies, between wells is difficult because of the distance between wells. However, Enclosure 2 illustrates the continuity between some parts of the Late Miocene-Pleistocene succession, as well as highlighting lateral variation and regions within the study area that appear disconnected from other regions.

4.5.1 Depositional model for northern Taranaki Basin

Posamentier et al. (1988a, p. 110; after Brown and Fisher, 1977) define a depositional system as a 'three-dimensional assemblage of lithofacies, genetically linked by active (modern) or inferred (ancient) processes and environments'. Specific wireline log motifs indicate types of facies that can be related to these depositional environments, though no pattern is unique to a particular environment (e.g., similar stacking patterns can be seen in fluvial, deltaic and deep marine environments; Cant, 1992). Delineation between depositional environments is therefore dependent upon association with other facies, and known (or inferred) position within the shelf-slope-basin. Biostratigraphic reconstruction of paleoenvironments (Chapter 6), in association with seismic reflection profiles (Chapter 5) place the offshore northern Taranaki Basin in a dominantly bathyal environment through the Miocene, shallowing up to shelfal depths during the Late Pliocene and Pleistocene. Consequently, this study can disregard deltaic or fluvial depositional environments, and instead place wireline facies almost entirely within slope to basin floor (to outer/mid shelf at more southerly and easterly sites) depositional settings.

Lateral and vertical variations in log pattern are often subtle and thereby difficult to interpret in terms of changes in the depositional environment (Danielsen et al., 1997). This is particularly true of the Giant Foresets Formation, where the prevalence of muddy lithologies mask or subdue wireline log motifs that may otherwise be more indicative of a particular depositional system or environment. Lack of well-defined biostratigraphic control at many sites, combined with the wide spacing of boreholes, only serves to aggravate this problem. It is anticipated that correlation with seismic reflection profiles will resolve this.

Regardless of the above-mentioned problems, however, wireline facies interpretation has served to delineate a number of distinct depositional systems. Throughout the Miocene, sedimentation in the study area was dominated by hemipelagic deposition as a direct result of the bathyal water depths that pervaded, and the distance from source. Coarse clastic material was only deposited as a result of intrabasin volcanic activity, although in the Late Miocene there was some interfingering of volcaniclastic facies with southerly-derived siliciclastic sandstone deposited ahead of the progradational wedge along the eastern edge of northern Taranaki Basin.

The latest Miocene to mid or Late Pliocene was a time of sediment starvation across much of the study area, resulting in deposition of extremely condensed calcareous units and the formation of an almost region-wide paraconformity. Only in the central part of the Northern Graben did active sedimentation occur, with deposition of terrigenous material associated with an aerially restricted basin floor fan system. Sedimentation during the Waipipian to Early Nukumaruan was dominated by hemipelagic and silty to sandy turbiditic sedimentation ahead of the advancing foreset front. Coarsening of facies units, and at the most northern and western well sites, collapse and slumping indicate arrival of this front at each site. Shelfal facies dominate the sedimentary sequence after the prograding front past a particular well site.

4.6 Cyclostratigraphy through the Giant Foresets Formation – is repetition evident on well logs?

4.6.1 Systems tracts concepts

The recognition of cyclical repetition of lithologies in a sedimentary record is a key to understanding sea level changes, particularly the influence of tectonic versus eustatic controls on sequence architecture. The sequence stratigraphic model of Vail (1987), Posamentier and Vail (1988), Van Wagoner et al. (1988), and Weimer and Posamentier (1993), among others, typically associates systems tracts with particular segments on the relative sea-level curve (Posamentier et al., 1988a). A systems tract is defined as a linkage of contemporaneous depositional systems (Van Wagoner et al., 1988, after Brown and Fisher, 1977), defined objectively by stratal geometries at bounding surfaces, position within the sequence, and internal parasequence stacking patterns (Miall, 1997).

Lowstand systems tracts (LST) are deposited basinward of the preceding depositional shoreline break during the lowest part of a sea level cycle. The boundary at the base of a LST is a sequence boundary. Transgressive (TST), highstand (HST), and regressive systems tracts (RST) are deposited during the rapidly rising, highstand, and falling limbs of relative sea level cycles, respectively. Regressive systems tract (RST) is a more recent addition to systems tracts terminology (see e.g., Hunt and Tucker, 1992, 1995; Posamentier et al., 1992; Kolla et al., 1995; Naish and Kamp, 1997), and have led to disagreement regarding the placement of the sequence boundary relative to this systems tract (e.g., Hunt and Tucker, 1992, 1995; Helland-Hansen and Gjelberg, 1994). Naish and Kamp (1998) showed that the sequence boundary lies above the

RST. Forced regressive systems tracts (FRST) are identified where a regressive surface of erosion (RSE) lies between the HST and the repressive deposits. Highstand systems tracts (HST) are associated with the relative sea level highstand. The underlying surface is known as the downlap surface, which approximates the maximum flooding surface (a conceptual surface). This latter systems tract becomes conformable on the inner shelf and loses its lithological identity within coastal plain sediments. The seismic character of each of these systems tracts are discussed further in Chapter 5. A more in-depth synopsis of systems tracts concepts, with figures, is given in Appendix 5b.

4.6.2 Sequence architecture

Kamp et al. (2002 and in prep.) and Vonk et al. (2002) have demonstrated that sequences, comprising RST, HST, and TST components, can be readily interpreted from wireline and outcrop investigation of shelfal sediments in the Wanganui Basin (particularly the Matemateaonga Formation). Delineation of each cycle is dependent on the identification of discrete shell beds, which display low GR, high resistivity and high sonic spikes on wireline logs (Fig. 4.23). These shell beds represent transgressive deposits, occurring as either onlapping, reworked shell lags (Type A shell bed of Abbott and Carter, 1997) or backlapping *in situ* shell beds which accumulated at inner to mid shelf depths (Type B shell bed of Abbott and Carter, 1997).

The characteristic succession of shell bed (TST) -siltstone (HST) -sandstone (RST) lithofacies is documented in Kamp and McIntyre (1998) and Naish and Kamp (1997), with each association typically in the order of 25-70 m thickness. Sequences identified in this manner are interpreted to have a periodicity of 100 or 41 ka (5^{th} or 6^{th}-order) in Late Miocene through Pleistocene strata. Aside from outcrop that can be keyed into nearby wells, Wanganui Basin studies have an advantage in that a continuously cored well (Manutahi-1; Robinson et al., 1987) provided an excellent opportunity to compare the lithofacies of shelfal systems tracts of the Matemateaonga Formation with their geophysical wireline log character. This has enabled shelfal Matemateaonga Formation deposits to be widely mapped in the subsurface of Wanganui Basin and Taranaki Peninsula on the basis of specific wireline responses to particular lithologies (Griffin, 2001; Kamp et al., 2002).

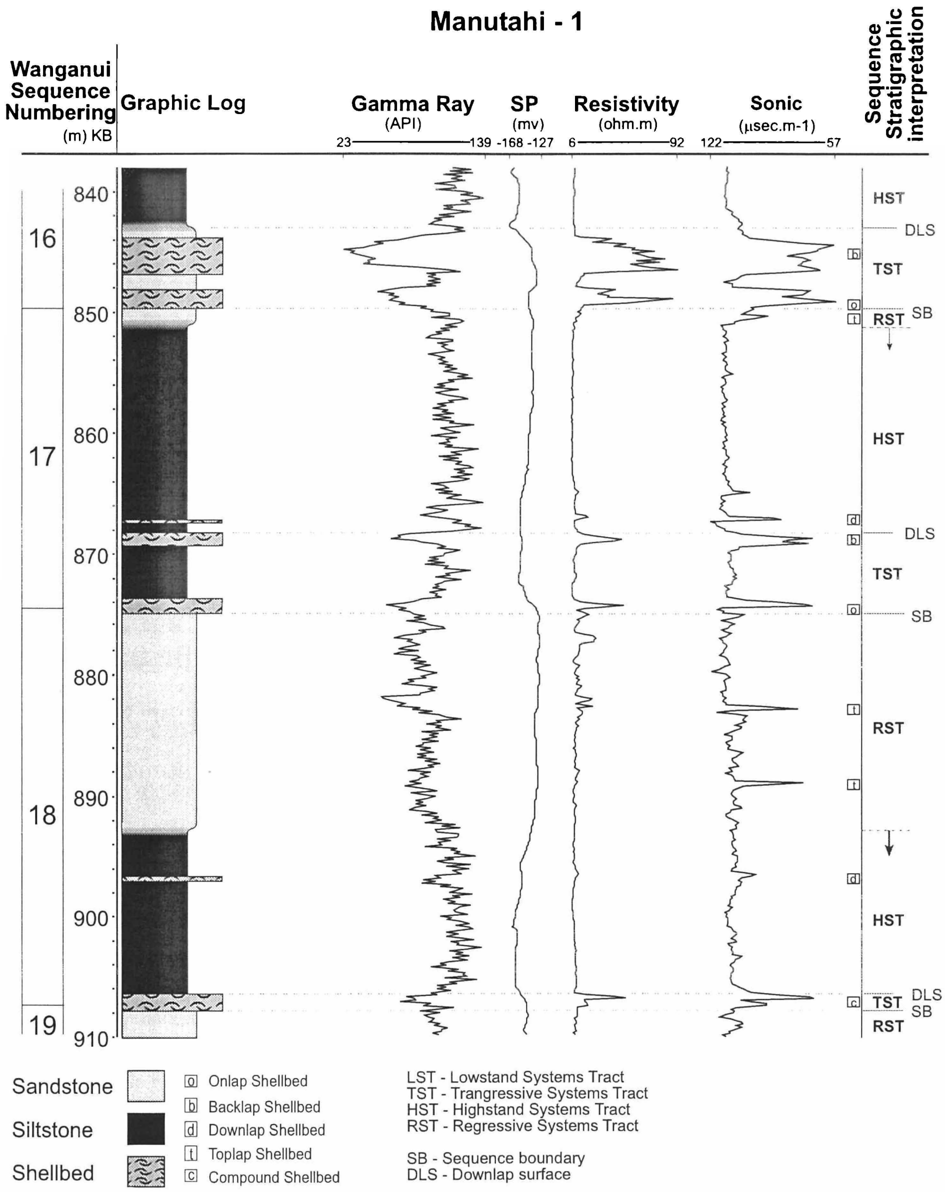


Fig. 4.23: Stratigraphic column, wireline logs, and sequence architecture of part of the Matemateaonga Formation, Manutahi-1. Each cycle is delineated by the presence of shell beds, identified from low GR, and high resistivity and sonic spikes. In the offshore northern Taranaki Basin region, depth of deposition (predominantly bathyal) precludes the identification of cycles on the same basis (compare with Fig. 4.23). Log from Kamp et al. (2002). See reference for explanation of terminology.

However, while sequences can be easily defined in outcrop and in wireline logs through the Matemateaonga Formation, this is not the case in the shelf deposits (topsets) of northern Taranaki Basin. This is because shell beds are not well developed and the lithologies are mud dominated and seem to lack the thick sandstone beds of RSTs. In addition, wireline logs are not available for the upper 400-500 m of shelf (topsets) strata.

These problems are highlighted by Fig. 4.24 - compare, for example, the wireline motifs displayed at this site through the Giant Foresets Formation, with the wireline motifs through the Matemateaonga Formation in the Manutahi-1 log (Fig. 4.23). Figure 4.24, which illustrates wireline trends through the topsets (or shelf deposits) of Kora-1, shows that, while the GR log clearly displays cyclical patterns (cyclothems), these trends are not mimicked by the other logs. While these trends *do* exist in the shelfal sediments (substantiated by textural data), and actually display a positive correlation with seismic reflectors (Chapter 5), this pattern can be replicated in few wells; the interval of shelf deposits logged at each site is not enough to draw any definitive conclusions on the nature and periodicity of cyclicity through shelf sequences.

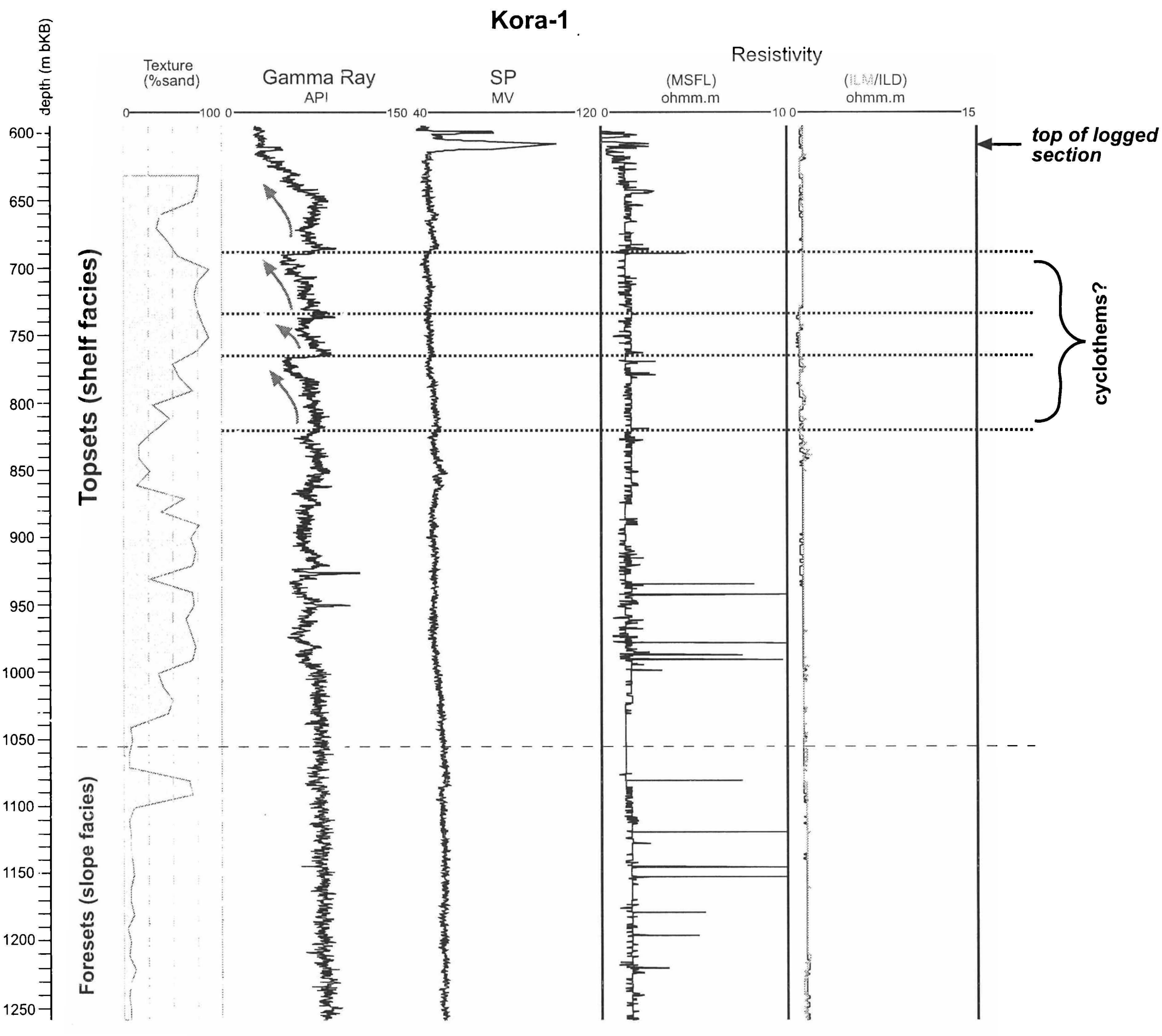


Fig. 4.24: Identification of cyclical packages within the Plio-Pleistocene succession is hampered by often featureless wireline log motifs, and lack of logged section in the upper part of holes. This example (Kora-1) illustrates the difficulty and ambiguity associated with identification of higher-order cyclical packages in the upper part of the Giant Foresets Formation. The sequences marked seem to be reflecting textural change, but lack the shell beds that give rise to spikes in resistivity and sonic velocity.

Basin floor fan sandstone beds can be used as a proxy for sea level change, as coarser sediment is more likely to be deposited during periods of lowering or low sea level (LST; e.g., King et al., 1994). However, delineation of cycles on this basis is only possible where obvious basin-floor fan facies can be identified on geophysical logs and corroborated by well cuttings, such as the basin floor fan facies of the Mangaa Formation (Fig. 4.25). Deposition of the Mangaa Formation was confined by the Northern Graben, which may have influenced formation of part of the depositional motif, and may therefore not be due entirely to sequence stratigraphic controls. The bottomsets and slopesets of the Giant Foresets Formation, as imaged on wireline logs, do not seem to show the depositional characteristics of Mount Messenger Formation and Urenui Formation basin floor and slope fans, which could be regarded as outcrop analogues. The predominantly muddy lithologies of the Giant Foresets Formation may indicate sediment by-pass, at least in the areas in which wells are sited, much like that postulated for the Urenui Formation (King et al., 1993, 1994).

While the above discussion emphasises the difficulty of recognising sequences through the Giant Foresets, it is not entirely impossible to do so. In fact, recognition of LST prograding complexes (channel levees, slope fans etc.), and slumped facies can help to delineate a number of cyclical packages, together with decreasing or increasing GR/SP trends. This has enabled the various components of the lowstand prograding complex to be characterised on the basis of their wireline motifs. Well log patterns in the prograding complex have a variety of expressions, depending where they occur within the complex (e.g., channel, levee, overbank deposit), but the most common trend displayed is the upward-coarsening motif, associated with an increasing grain size or decrease in the clay/silt content (Sf1). This reflects higher rates of sediment accumulation relative to accommodation, resulting in basinward migration of depositional environments to produce a relative shallowing upwards or regressional sequence (Mitchum et al., 1993; Ahmadi and Coe, 1998). In slope and basinal positions that are less sand rich (e.g., lower part of the Giant Foresets Formation), well logs typically lack the diagnostic coarsening-upward patterns.

Conversely, HST (or retrogradational phase) deposits are characterised by upward-increasing GR values, reflecting the lower energy conditions associated with higher relative sea levels, and a corresponding increase in finer-grained sediment. Highstand deposits ('dirtying-upward' trend) may record the waning of a depositional system, and implies significant deepening that occurs when sediment supply is less than the rate of creation of accommodation volume (Mitchum et al., 1993).

Examination of the Giant Foresets Formation reveals that several discrete successions of stacked sequences (predominantly Sf1 and Sf2 successions) displaying an overall coarsening-upward trend, can be identified on GR logs (Figs. 4.25, 4.26). These progradational trends are subtle, and are often overprinted in the lower part of the Giant Foresets Formation by a more aggradational pattern, characterised by an erratic and irregular GR log response. Aggradational patterns are attributed to the rate of accommodation and volume of sediment supply being roughly equal, resulting in vertical stacking of facies (Mitchum et al., 1993). This suggests that, during the Pliocene, sediment supply and creation of accommodation space were roughly in balance, before the former outstripped the latter as a result of large volumes of sediment being introduced to the basin from erosion of the Southern Alps. Higher in the sequence, slumped facies are indicative of lowering sea level conditions, and are placed in the LST.

While progradational trends and associated stacking patterns are more commonly identified, at Wainui-1 (Appendix 2) fining-upward sequences probably indicative of increasing water depths (TST-HST deposits?) are identified. King et al. (1994) attribute similar successions in the upper Mount Messenger Formation, which display an overall fining-upward trend, to allocyclic controls and relative base level rise. Why this site displays these retrogradational patterns when other sites do not is debatable. It may simply be due to the distal locality of this site, always being at bathyal depths (see Chapter 6), and the most basinward of any site studied. However, it is more probable that proximity to large feeder channel systems (Fig. 2.5) has resulted in the trends being displayed at this site. During lowstands these channels would have acted as conduits for sediment transportation, disgorging coarser material as distal (basin floor) fans and overbank deposits. A rise in relative sea level would have progressively shut off this depositional system, recorded by the upwards-fining retrogradational succession. HST deposits are difficult to interpret from wireline log motifs, particularly through the predominantly muddy Giant Foresets sequence.

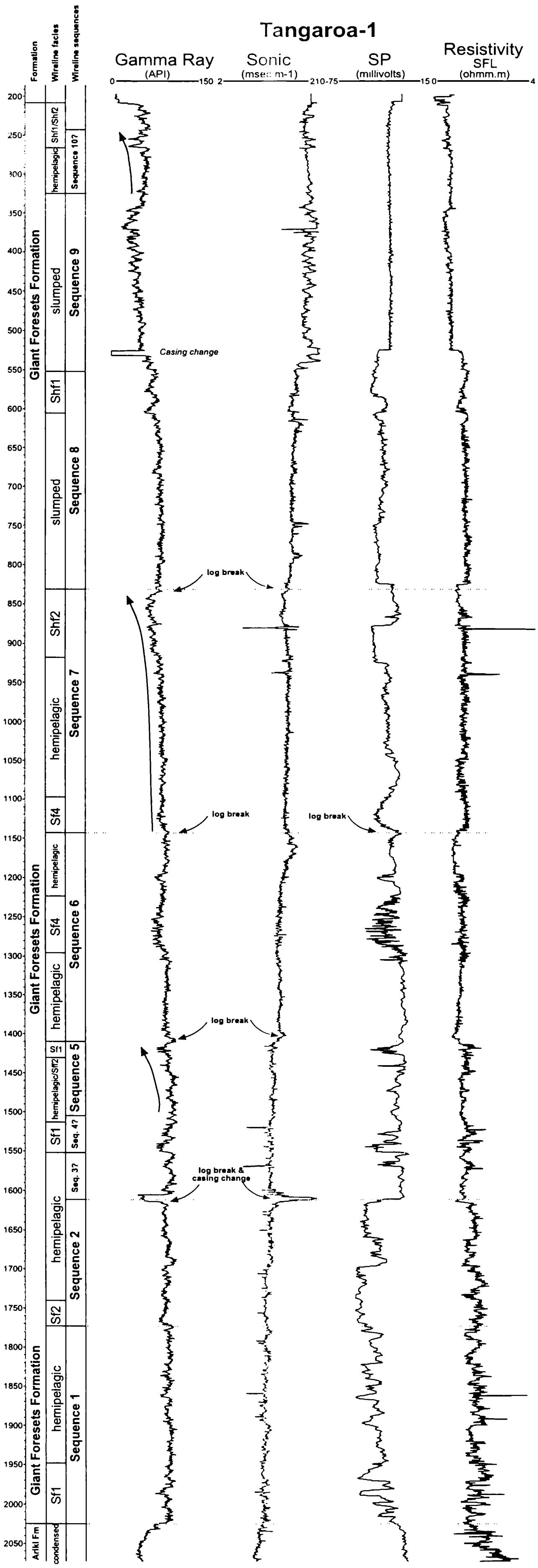


Fig. 4.26: Up to 10 cycles, based on wireline interpretation, can be identified through the Giant Foresets Formation at Tangaroa-1. Each cycle is delineated by an overall coarsening-upward pattern, either of the GR/sonic or SP/resistivity logs, or by a distinct break in all logs. Where slumps are present, the basal contact is taken as the sequence boundary, otherwise the boundary is marked by the coarsest unit at the top of the upward-coarsening package.

Even in analogous outcrop sections (upper Mount Messenger Formation), King et al. (1994) can only postulate that mudstone overlying thin-bedded sandstone units could be attributed to distal overbank deposits of a channel-levee complex, toe deposits of a prograding complex, or pelagic highstand systems tract deposits, as the Vail model places most of the HST on the shelf. The thick mudstone units separating the lower (Mangaa B sands) and upper (Mangaa A sands) basin floor sandstone beds of the Mangaa Formation, and indeed the blocky sandstone units within each stacked succession, similarly may be inter-fan deposits related to lobe switching, or may be due to background sedimentation during highstand conditions.

Within the Plio-Pleistocene strata of northern Taranaki Basin, the sequence architecture of any one sequence is dependent on the location of the site. For example, lower sequences at Mangaa-1 and Awatea-1 comprise RST/LST basin floor fans, and may or may not incorporate HST deposits. Higher in the succession, sequences are dominated by progradational GR log trends which attest to increasingly higher energy environments, but cannot be confidently subdivided into systems tracts components (other than being attributed to the RST/LST prograding complex). Tangaroa-1 displays a number (up to ten?) of coarsening-upwards progradational successions (Fig. 4.26), but at other sites it is difficult to identify any (e.g., Kahawai-1), and at Wainui-1, retrogradational successions are more common. Because of this, correlation of these cyclical trends is not yet attempted (see Chapter 5).

4.6.3 The sequence boundary

In the sequence stratigraphic models of Vail (1987) and Posamentier et al. (1988), the sequence boundary is placed below the massive blocky sands of the basin floor fans. Other models placed the boundary above the fans (e.g., Hunt and Tucker, 1992, 1993, 1995; Helland-Hansen and Gjelberg, 1994) arguing that the sequence boundary should be most extensively developed at the lowest point in the sea level cycle. It follows that, as basin floor fans predominantly accumulate during *lowering* sea levels, the sequence boundary should thus be placed at the top of these deposits (RSTs). Where slumped horizons are present (presumed to be deposited at a later stage than LST sandstone beds), the sequence boundary is then taken as the erosional basal contact (see Hansen, 1996). While King et al. (1993, 1994) adhere to the classical Vail model, the alternative of having the sequence boundary as a correlative conformity at the top of a basin floor succession is followed in this study. In addition, the top of each wireline sequence (top of the coarsening-upwards succession) is often delineated by a defined log break (Fig. 4.26), which may be indicative of a thin starved condensed interval (hemipelagic TST/HST deposits; Mitchum et al., 1993). Further examination of the nature of the sequence boundaries follows in Chapters 5 and 6.

4.6.4 Order of cyclicity

Overall, only subtle changes in lithology occur upward through the Giant Foresets Formation, which makes it much more difficult to construct a record of higher-order sequences on the basis of wireline motif, and almost impossible to accurately interpret the various components of a relative sea level cycle. However, at least three orders of cyclicity can be recognised:

1. the tectonically controlled 2nd-order cycles identified by Kamp et al. (2002), including the Whangamomona Group and the Rangitikei Supergroup (which includes the Giant Foresets Formation in their nomenclature);
2. progradational or retrogradational successions that comprise RST/LST and possibly TST/HST deposits (4th-order);
3. finer-scale upward-coarsening units, which comprise the 4th-order progradational or retrogradational successions, and sandstone-mudstone couplets within stacked basin floor fan facies (5th-order?).

Tangaroa-1 (Fig. 4.26) displays a more well-defined series of progradational sequences compared with any other site in the northern part of the basin, and contains at least 10 wireline log defined sequences. This equates to roughly 3.5 m.y. of sedimentation, with an approximate cyclicity of about 350-400 ka (4th-order). As illustrated in Figs. 4.24 and 4.25, the upper part of the Giant Foresets Formation often contains numerous stacked cyclothemic packages, often grading up from claystone to siltstone (Sff2) or from claystone to siltstone to sandstone (Sff1). At Mangaa-1, these packages are often capped by a shelly horizon. This suggests an order of cyclicity approaching that of Mid to Late Pleistocene shelfal deposits of the Wanganui Basin (possibly 5th –order or 100 ka periodicity). However, as discussed in the previous section, the lack of defining and consistent criteria for the identification of sequences means that delineation of sequences is arbitrary using wireline log data alone. While these cyclothemic packages may indeed be the product of eustatically-controlled sea level changes, and may occur in the order of ?100 ka, they may also be the product of a series of closely-spaced catastrophic and instantaneous 'events', e.g., the products of several turbidity currents and other mass flow mechanisms arising from autocyclic controls.

4.7 Conclusions

Use of geophysical wireline logs, in conjunction with lithological logs obtained during the drilling process, and biostratigraphic age-datums, is an effective way of correlating formations in the subcrop. By correlating Wanganui and Taranaki Basin strata, it is possible to understand more clearly the link between the two basins.

- Sedimentation patterns in northern (offshore) Taranaki Basin are intimately linked to contemporaneous deposition in Wanganui and King Country Basins. The continental margin associated with the Whangamomona Group only had limited progradation into Taranaki Basin during the Middle Miocene to Early Pliocene, evidenced by thick accumulations of basin floor and slope deposits along the eastern margin of the basin (northern Taranaki coastal outcrop and nearshore wells), and the predominance of hemipelagic sedimentation elsewhere in northern Taranaki Basin.
- The Ariki Formation is a significant marker horizon in offshore northern Taranaki Basin. Based on its wireline motif, time-equivalents of this formation may be more widespread than previously recognised. This formation is particularly relevant as it is related to at least three important events: (i) the limited extent of progradation of the Whangamomona Group foreset front, (ii) initiation of extension of the Northern Graben, and (iii) the timing of the Tangahoe pulldown in Wanganui Basin.
- The wireline motif of the Mangaa Formation is, like the lower part of the Mount Messenger Formation, characteristic of basin floor fan sedimentation. However, the Mangaa Formation appears to be extremely restricted in its extent, indicating that the Northern Graben acted as a sink for sediment, probably initially sourced from contemporaneous uplift to the east.
- It is difficult to characterise the Giant Foresets Formation on the basis of geophysical wireline motif. Often only very subtle changes (in wireline motif) are noted uphole, although GR (and occasionally sonic and density logs) tend to display an overall coarsening-upward trend. At most sites, the upper few hundred metres of each logged well section are distinctly different, reflecting the coarser deposits of the shelf facies (topsets).

- Six facies based on wireline log motifs have been mapped in the northern Taranaki Basin: (i) hemipelagic facies, (ii) basin floor fan facies, (iii) slope fan facies, (iv) shelf facies, (v) slumped facies, and (vi) condensed facies.
- The Miocene is dominated by hemipelagic (basinal) facies, with interfingering of coarser lithologies of the volcaniclastic basin floor fan facies (Bff2), and volcaniclastic Sf3 sediments in the north and east of the study area.
- Pliocene basin facies (bottomsets) are also dominated by hemipelagic sedimentation (other than the Mangaa Formation), although both the sporadic occurrence of thin basin floor fan facies and the increasing prevalence of Sf2 (distal fan sedimentation?) are suggestive of the increasing proximity of the advancing foreset front.
- Arrival of the foreset front at any one site is documented by the frequent presence of Sf1 and slumped facies. The latter facies are more often than not associated with degradational foresets, and a change in the physical properties of the clay matrix.
- While wireline logs are an effective tool for identifying depositional environments, and in shelfal settings can be used effectively for identifying sequences and systems tracts, serious limitations arise in deeper water situations. Within offshore northern Taranaki Basin sediments, bathyal water depths, which dominated until the Pleistocene, and distance from the depositional shoreline, resulted in a predominance of muddy hemipelagic facies. Though the Giant Foresets Formation does display fine scale coarsening- or fining-upward cyclical packages (Sf1, Sf2, and Shf1), resolution of the logging tools, and lack of well-defined biostratigraphic dating, means that ambiguity is inherent in any interpretation.
- Three levels of cyclicity can be recognised: 2nd-order (several millions of years) related to progradation of the Whangamomona Group and Rangitikei Supergroup, 4th-order (400 ka) cycles, and ?5th-order (100 ka) cycles (possibly related to glacio-eustatic controls). Neither the 4th or 5th-order cycles can be consistently correlated between well sites.

Chapter 5: Seismic geometry and sequence architecture of the Giant Foresets Formation, with emphasis on the Northern Graben region

Chapter 5: Seismic geometry and sequence architecture of the Giant Foresets Formation, with emphasis on the Northern Graben region

5.1 Introduction

Seismic stratigraphy is an integral part of this study, particularly in light of the fact that the Giant Foresets Formation takes its name from the characteristic sigmoidal-shaped off-lapping clinoform appearance displayed on seismic reflection profiles (Shell BP Todd, 1976). Interpretation of seismic reflection profiles is fundamental to understanding several aspects of the general evolution of the Giant Foresets Formation, and indeed the northern offshore Taranaki region. These include the extent of shelf progradation through time, the availability and creation of accommodation space, the stratal architecture of the succession, as well as definition of the intervals of major faulting, identification of the sediment pathways and depositional mechanisms, and the effect of topographic and structural features on sediment distribution. Because seismic reflection horizons are considered to be isochronous surfaces, they also provide a method of correlating biostratigraphic datums (as outlined in Chapter 3) between well sites.

A large amount of industry-acquired multi-channel seismic reflection data over the study area are now open file. This study has utilised some of the most recent open file data from PR 2261 (Petrocorp, 1995), focusing on the Northern Graben, as well as several older lines over the less structurally complex Western Stable Platform. The quality of data varies from 24-fold migrated data for the older lines, to 60-fold migrated data for the more recently acquired data. Appendix 5a lists the seismic reflection data used in this study, as well as the co-ordinates (New Zealand Map Grid) for the start and end points of each line.

5.1.1 Basic principles of seismic stratigraphy

Seismic reflections are generated along surfaces of acoustic impedance contrast, a bulk physical property of a rock and its contained fluids, and are obtained by multiplying the density of a rock by the velocity of acoustic waves passing through it (Cross and Lessenger, 1988). The application of seismic stratigraphic concepts has grown markedly since the 1970's (e.g., Brown and Fisher, 1977, 1984; Mitchum et al., 1977a,b; Posamentier and Vail, 1988; Posamentier et al., 1992) with the realisation that stratigraphic terminations, resulting from the difference in impedance between two rock layers, are often clearly imaged on seismic reflection profiles (see also Appendix 5b). These terminations are used to delineate well-defined seismic reflection packages or 'sequences', which represent genetically related, depositional (stratigraphic) units, bound by unconformities and their correlative conformities (Mitchum et al., 1977b; Van Wagoner et al., 1987). Primary depositional sequences can often be divided into systems tracts. Systems tracts are linked depositional systems related to a particular segment of a related relative sea level curve (Brown and Fisher, 1980).

Seismic facies units (architectural elements), interpreted on the basis of differences in internal reflection configurations from those of adjacent facies units (Mitchum et al., 1977a), are used to interpret depositional processes and environments and possible sedimentary lithofacies. Because seismic reflections are often isochronous, they may pass laterally through lithostratigraphic facies boundaries. For example, in the model presented in Fig. 2 (Appendix 5b), a single reflector horizon may pass from coarse-grained nearshore sands through to muddy slope sediments, then to sandy basin-floor fan sediments. Within sequences, seismic facies may terminate abruptly against other seismic facies units, but may also grade into other facies. Boundaries of seismic facies may be sharp and easily identified, or may be transitional (Brown and Fisher, 1980; Vail, 1987).

Seismic sequence stratigraphy is a common methodolgy utilised in the petroleum industry to understand depositional environments and to predict lithofacies. However, it is recognised that not all seismic successions can be satisfactorily interpreted using accepted recognition criteria. Such is the case with seismic reflection profiles used in this study. Often, it is difficult to recognise sequence-defining onlap reflection configurations or terminations, and few seismic units display the 'typical' components of a seismic sequence as defined in the classic Exxon-type model (Appendix 5b). Furthermore, the dominantly muddy nature of the Giant Foresets Formation and scarcity of well sites in the study area precludes the confident recognition of particular sequence components through facies change and/or wireline log motifs. Thus, this study has taken more of a stratigraphic approach, describing the gross depositional (seismic) characteristics of each of the major formations investigated in this study (section 5.4). This has been followed by more detailed mapping of the geometry and distribution of discrete seismic units within the Giant Foresets Formation (section 5.5), and description of the seismic facies (or architectural elements) that can be recognised within these seismic units, with interpretation of their depositional mechanisms and depositional environments (section 5.6). Section 5.8

discusses architectural elements in the context of their relationship to sea level change, sediment supply and type, and places these elements into sequence stratigraphic models where applicable.

5.2 Previous seismic studies

While the Giant Foresets Formation is imaged exceptionally well on seismic reflection profiles, appearing as a succession of stacked and progressively offlapping sigmoidal-shaped clinoforms (King and Thrasher, 1996), few detailed studies have focused on the detailed geometry and architecture of this formation. Thrasher and Cahill (1990) initially mapped the thickness and distribution of post-Miocene strata, using seismic reflection profiles available at the time, while Beggs (1990) divided the clinoform units that comprise the Giant Foresets Formation into four discrete components (after Mitchum et al., 1977a; Sangree and Widmier, 1979; Fig. 5.1) based on position within the shelf-slope profile, and seismic expression:

1. *Topsets* - Characterised by subparallel, subhorizontal, and moderately continuous, moderate to high amplitude reflectors. Lithologies include sandstone, muddy siltstone, and shell beds or disseminated shell hash, consistent with accumulation on a continental shelf.
2. *Progradational Foresets* – The seismic reflectors of these slope facies are coherent and off-lapping, dipping basinward at 1-3°, are moderately continuous, with moderate amplitude. Lithologically, these foresets are dominantly fine-grained, and are comprised of monotonous mudstone and muddy siltstone.
3. *Degradational Foresets* - These reflectors represent deposition on a continental slope also, but the reflectors dip at steeper angles than above (up to 10°), and are of lower amplitude. Internally, they are more chaotic, reflecting mass movement downslope. The units are lithologically variable, consisting of mudstone (dominant), with sporadic sandstone and conglomeratic beds.
4. *Bottomsets* – These sub-horizontal to slightly inclined reflectors represent deposition on a basin-floor. Reflectors display variable continuity and are generally moderate to low-amplitude. Lithologically, units ascribed to this division are variable, including sandstone and mudstone.

Later studies by Soenander (1991, 1992), focusing on the Giant Foresets Formation in the vicinity of Taimana-1 and Arawa-1, used seismic stratigraphic concepts to map four major depositional units (Soenanders' units A-D), with each of these units separated by a prominent reflector, characterised by termination of the internal reflectors against this reflector. Isopach and structure-contour mapping illustrated the direction of progradation of the foresets, and demonstrated that the clinoforms on the Western Stable Platform prograded in the form of migrating fan lobes. Soenander (1991) recognised only three divisions based on their seismic character, excluding the degradational foresets group, as these are not present in the vicinity of his study area. Previous studies in the southern part of Taranaki Basin have determined sedimentation rates and linked a sudden increase in the rate of Pliocene-Pleistocene outbuilding of the continental shelf to uplift of the Southern Alps (Bosario, 1981; Ogilvie, 1993), and discussed the influence of eustasy, subsidence and sedimentation on the distribution of sediments and the nature of unconformities (Ogilvie, 1993). Other studies (e.g., Hayward, 1990; Crundwell et al., 1994) have used foreset sequences to calibrate the paleobathymetric position of certain depth-restricted benthic foraminifera by extrapolating present (water) depths to modern topsets, foresets etc., and estimating the depth to a point in the ancient record (see also Chapter 6).

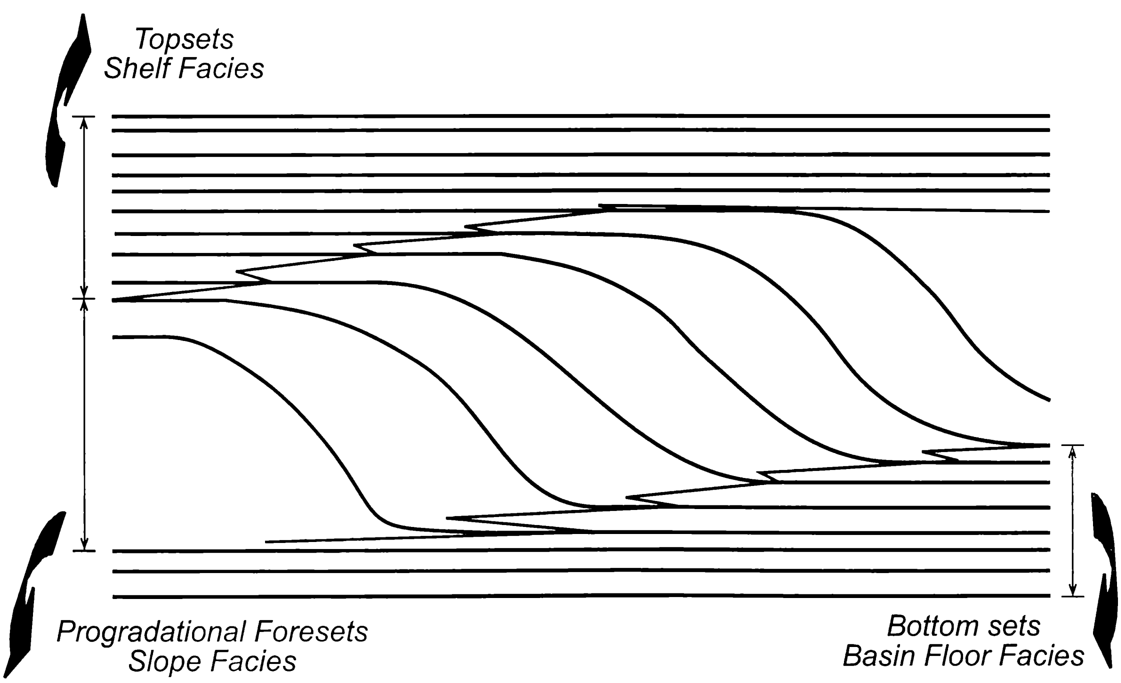


Fig. 5.1: Schematic representation of the foreset divisions of Beggs (1990). See text for seismic characteristics.

This chapter builds on these previous studies, and is intended to provide a broad seismic framework on which subsequent studies can be further developed.

5.3 Methods of analysis and interpretation

5.3.1 Data sets

Thirty-two seismic lines, including 25 from the P95-series (Petrocorp Exploration, PR 2261, 1995) and seven from various earlier data sets (Appendix 5a), have been interpreted during the course of this investigation. The data set includes 11 strike lines (seismic surveys which run parallel to the associated shelf edge), 18 dip lines (seismic surveys that run perpendicular to the associated shelf edge), and three lines running at an oblique angle to the shelf edge (Fig. 5.2). Initial interpretation of all seismic lines was achieved using paper records, obtained from the Ministry of Economic Development (Wellington). The P95-series of seismic reflection profiles, shot in 1995, are of exceptionally good quality, and are far superior in terms of resolution of seismic horizons than obtained in previous seismic studies. Because the Western Stable Platform has been tectonically quiescent since the Miocene, only a minimal number of older lines (HF- and AR-series) were used to link key well sections in the western and southern parts of the Western Stable Platform (Wainui-1, Taimana-1 and Arawa-1) to the well sections to the north and east (Ariki-1, Awatea-1, Kahawai-1, Kora-1, Mangaa-1 Te Kumi-1, Tangaroa-1, and Turi-1). The positioning of seismic survey shot points is estimated to be accurate to within (+/-) 1 m for the P95 series of lines, and (+/-) 20 m for older shot points (Ministry of Economic Development, Wellington).

5.3.2 Outline of interpretation procedure

Seismic reflectors were initially mapped on paper traces using the loop-tie method, where all reflectors on all lines were tied back to previously interpreted lines, and to the primary dip line (P95-158). Seismic units were delineated on the basis of internal seismic reflection character and nature of the bounding reflectors (amplitude, brightness, continuity). Each unit identified was given a distinctive colour to distinguish it from units above and below. More than 90 discrete seismic units have been identified and mapped (each delineated by a letter or combination of letters; Appendix 5d), with many of these units able to be mapped over a wide area. Enclosure 3 displays examples of two uninterpreted and interpreted seismic reflection profiles from the northern part of the study area (P95-158, and intersecting strike line P95-163).

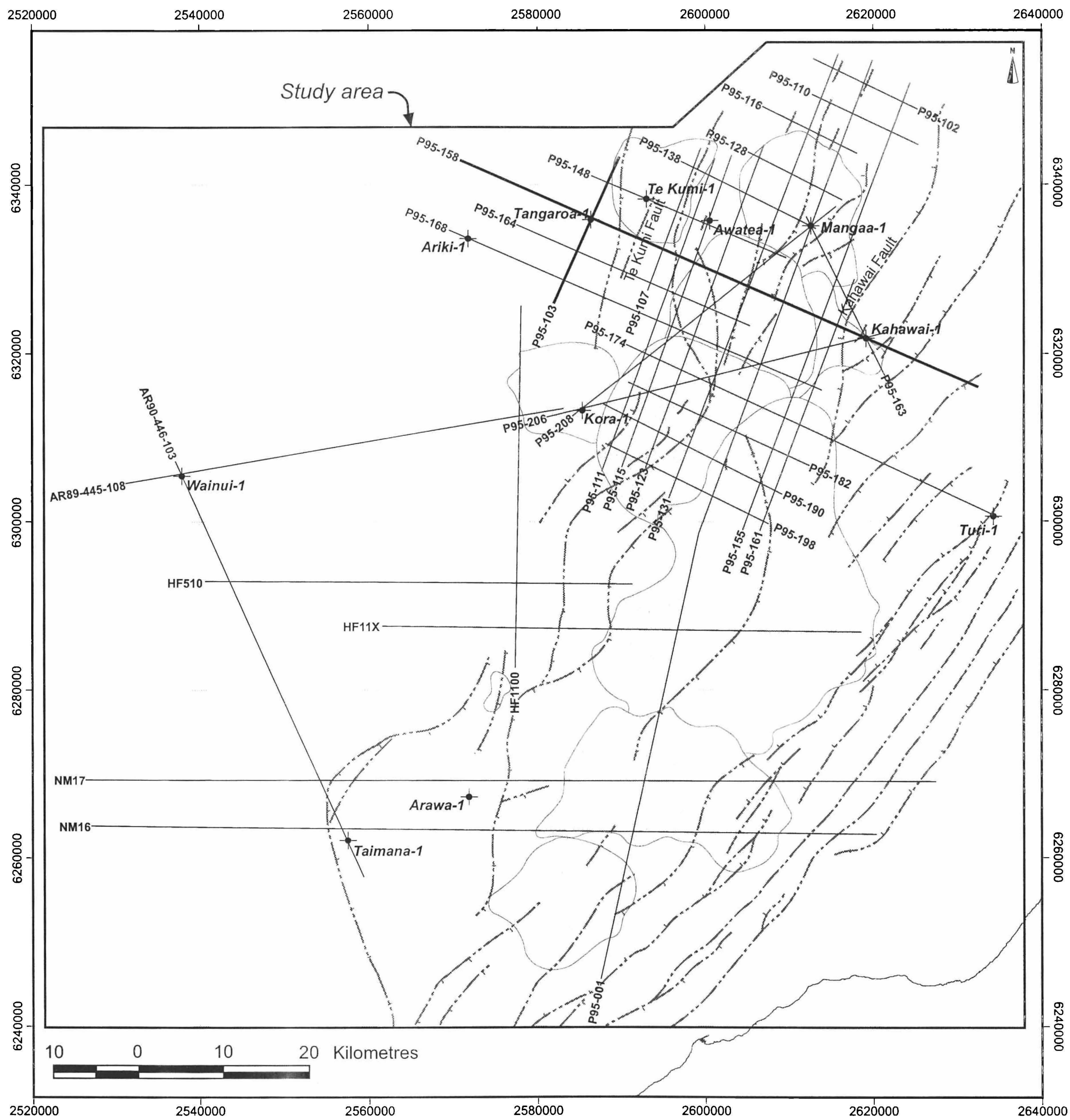


Fig. 5.2: Study area base map, illustrating location of seismic reflection profiles used in this study, as well as major faults (dashed lines) and volcanic massifs (grey areas; after Thrasher et al., 2002). Seismic reflection profiles P95-158 (primary dip line) and P93-103 (both bolded) are included on Enclosure 3.

5.3.3 Digitising and mapping of seismic reflection profiles

To aid in the mapping of seismic horizons, all lines in the P95-series were digitised using ArcEdit™, and values were interpolated for each horizon at specified intervals along the seismic line. Most lines were too long to be completely digitised at the same time, so were instead digitised in sections, and later joined and edited using ArcTools™. Once joined, a series of four short AML programmes (created by D. Gillgren, formerly of the Geography Department, University of Waikato, specifically for this study) were run on each line using ArcInfo™. Details of the digitising process and AML syntax are given in Appendix 6c.

The AML programmes created nodes at 500 m intervals along each seismic line, with a northing and easting (NZMS grid co-ordinates) attributed to each node. At each of these nodes, a number, in two-way travel time (TWT) was recorded for the top of every digitised layer intersected vertically below that point. This resulted in the generation of a spreadsheet that contained the depth (TWT) to each horizon at the specified interval distance (node), plus the grid co-ordinate for each node. Editing of spreadsheets involved checking generated data against the paper copy and assigning colour-coded names (according to the colours used on the paper traces) to the correct column. The thicknesses of individual seismic units in TWT were calculated by subtracting the depth for the layer below from the depth of the layer for which the thickness is calculated. When individual spreadsheets for each line had been completed, these were compiled into a composite spreadsheet for use in mapping in ArcMap™.

5.3.3 (i) Contour mapping

Structure-contour maps illustrated in this chapter were generated in ArcMap™. As the initial spreadsheet used is a composite of all P95-series lines, mapping was a simple process of selecting the column containing the related data for the colour-coded seismic unit and then using ArcMap™ to map out that unit. Two-way travel time structure-contour maps were contoured in 50 millisecond intervals. While in ArcMap™, a number of different 'themes' (characteristics), including the grid location of each seismic line, shot points, well locations, and coastline, were added. After generation of maps in ArcMap™, each was exported as a JPEG, and imported into the Macromedia Freehand 9.0/10.0™ drawing package, where contours were traced over and smoothed. Faults were added, and checked against the paper copy to see if they offset a particular horizon. Where throw on a fault was observed, contour lines were manually adjusted. Depths to relative horizons identified on the older HF series and AR series of lines were manually added at a later date.

A separate spreadsheet was created for generating structure-contour and isopach maps in metres. This spreadsheet was created using the velocity functions outlined in section 5.3.4, and maps were constructed in the same manner as described above. While an adequate number of seismic reflection profiles were interpreted over the Northern Graben, resulting in a grid of seismic lines approximately 2-5 km apart, wide spacing of lines over the Western Stable Platform suggest that errors in contouring are inherent. However, given the structural simplicity of this region, this is not considered to be a major concern as the resultant contours mimic the contour pattern mapped for the Miocene-Pliocene surface by Thrasher and Cahill (1990). All maps are uncorrected for compaction, isostasy, and subsidence.

5.3.3 (ii) Problems encountered

A major problem encountered during the digitising stage was that the computer processor was unable to run the AML programmes on long seismic lines, as it required too much computer memory to create the nodes and to give each node a value. To overcome this, each of the longer lines was split back into the smaller sections in which they were digitised. The four AML's were run on each section, and the resulting spreadsheets were manipulated and joined together.

Another problem encountered was somewhat of an enigma, and was one that could not be rectified in the AML's. Occasionally, large chunks of data at specific intervals were not generated. This was identified through manual checking of lines, and necessitated that all lines be checked. Depths, where missing, had to be manually added to the spreadsheet. While this took time, it was in fact an excellent check on the accuracy of the programme in assigning TWT values to digitised horizons (which was assessed to be fairly accurate).

5.3.4 Time-depth conversion

Time-depth data from wells Arawa-1, Ariki-1, Awatea-1, Kahawai-1, Kora-1, Mangaa-1, Taimana-1, Tangaroa-1, Te Kumi-1, Turi-1, and Wainui-1 were used to determine depths to formation tops and seismic unit tops at these sites. These data were supplied in the form of a time depth chart (Fig. 5.3), time-depth conversion formula (see below), and time-depth data spreadsheets, supplied by Geosphere Exploration Ltd. (Lower Hutt, New Zealand). As the spreadsheet data are of a commercially sensitive nature, the information from the spreadsheet is neither displayed in this chapter, nor in the appendices (other than Fig. 5.3).

Conversion of digitally-generated TWT data for each seismic horizon to depth in metres was achieved using the following sequential steps:

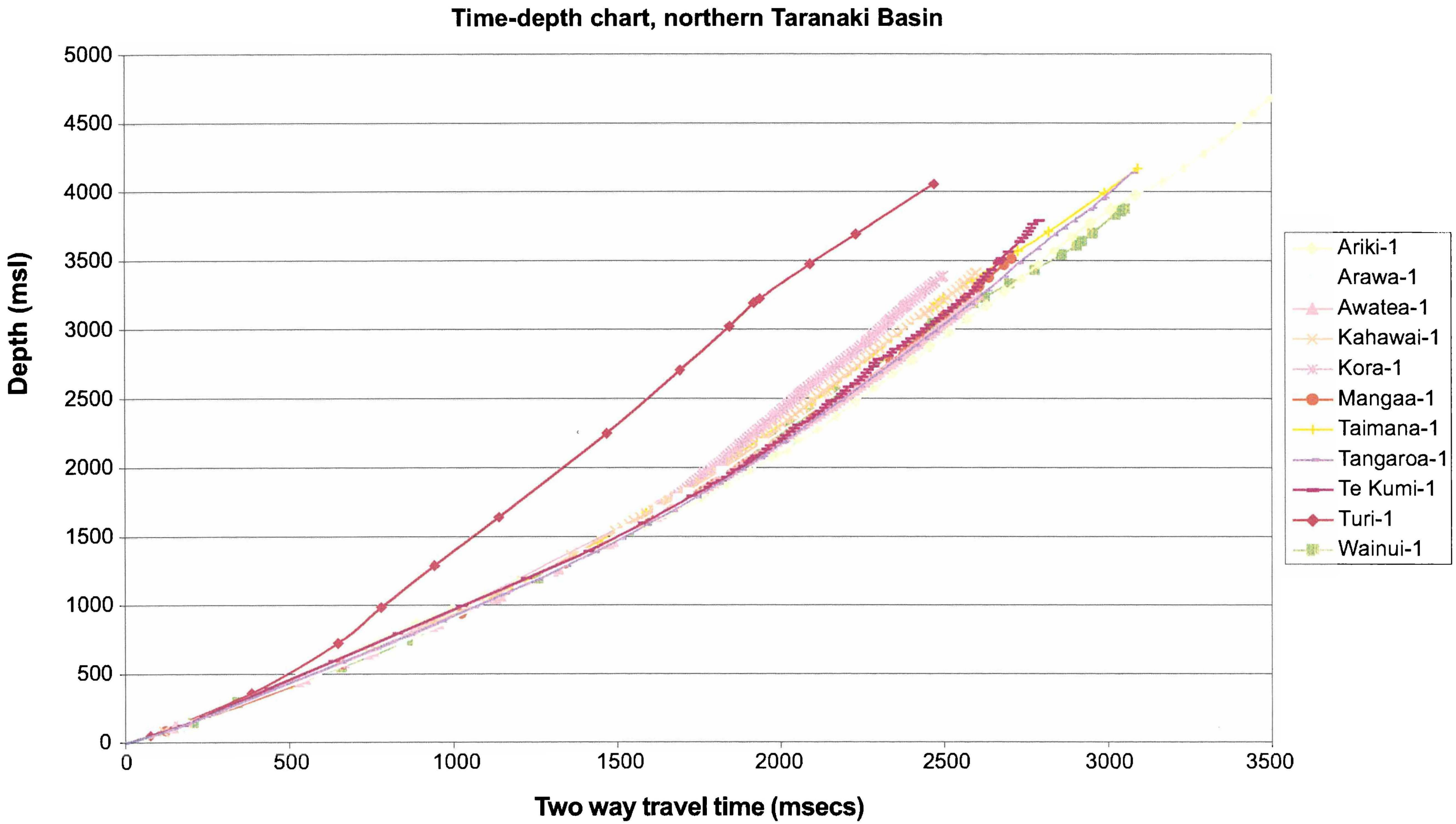


Fig. 5.3: Time-depth chart for northern Taranaki Basin wells. Proprietry data from Geosphere Exploration Ltd., Wellington. Note that Turi-1 lies off the typical trend of the time-depth data because of about 1200 m of uplift and erosion prior to the Pleistocene.

1. Depth from sea level to seabed was converted from time (in milliseconds) to depth (in metres) using a standard velocity function of 1500 m/second.
2. Depth from seabed to a specific horizon was calculated using the following formula:

$$h = ((1.7667\text{E-}4)\ (t\char`^2)) + (0.81911t)$$

Where:

h = thickness (from sea floor to horizon) in metres

t = two way interval time in milliseconds

3. Subsequent horizons were added together to get depth.
4. Thickness of each mapped seismic unit was calculated by subtracting the depth of the subsequent horizon from the depth of the next highest horizon.

The binomial supplied by Geosphere Exploration Ltd. was derived from a statistical fit to the checkshot data for 13 wells in the northern Taranaki Basin, and is only valid to a depth of about 2.5 km (G. Thrasher, pers. comm., 2001). While this equation has been calculated from large volumes of time-depth information for the northern part of Taranaki Basin, there will invariably be errors inherent in the calculation depending on depth and location, because of lateral changes in seismic velocity within each mapped unit. As well, the above equation does not hold true for submarine volcanic piles, nor is it valid for pre-Miocene sediments, or in areas where there are significant Neogene unconformities. Thus, the conversion factor does not hold for strata lower than the Miocene-Pliocene unconformity.

5.3.5 Sources of error

Stagpoole (1997) notes that acoustic scattering, diffraction and absorption within Miocene volcanic structures inhibits energy passing down into the region below, with seismic horizons often difficult to identify below these zones. This is also the case in this study area, but as this study is predominantly concerned with Late Miocene and younger sediments, the inability to accurately identify horizons below the many volcanic edifices in the region was not seen to be a significant problem. There is some distortion of reflectors over these massifs due to velocity pull-up (an effect caused by variation in lateral seismic velocities, which may artificially inflate doming over the structures e.g., Stagpoole, 1997) that may lead to interpretational errors.

However, the loop-tie method of seismic correlation used in this study has largely overcome this feature. Older lines, while still generally easy to interpret, do not display as high a resolution as the more recently acquired lines. Consequently, in many cases single reflector horizons on the older seismic profiles are in fact resolved into multiple reflectors on the more recent seismic profiles.

The use of different seismic surveys posed problems when tying in older surveys with more recent ones. At some tie-points, exact matches between reflector horizons could not be achieved (e.g., between AR90-445-103 and NM17). This problem is discussed in McQuillin et al. (1984), who indicate that while the data from any one survey will tie together more or less perfectly, lines from different surveys often will not. This is particularly true of older marine data sets, where less accurate position fixing can cause gross mis-ties, and differences in processing parameters can make the intersecting sections look very different from one another.

The scarcity and wide spacing of drill holes in the study area pose problems when trying to assess the continuity of formations. For example, parts of the Mangaa Formation, which is intersected in Awatea-1 and Mangaa-1, can actually be traced over a much wider area through reflection data, but is not identified on the basis of lithology in other wells even though correlative seismic horizons are present. The distance between wells means that lithological changes along a transect are not able to be assessed accurately, and facies relationships can only be estimated using a best-guess approximation based on current seismic stratigraphic models.

5.4 Seismic characteristics of underlying (Late Miocene) formations

5.4.1 Mohakatino Formation

On seismic reflection profiles, the volcaniclastic Mohakatino Formation is characterised by bold or bright, laterally continuous reflectors, forming bottomset (basin-floor) successions. Mohakatino Formation or equivalent sediments often pinch out up-dip against volcanic edifices, with individual units thickening away from the massif (Fig. 5.4). The boundaries of the volcaniclastic interval are also commonly delineated by bright, high amplitude horizons that display extensive lateral continuity. Internally, reflectors are sub-parallel, smooth to slightly hummocky and of variable amplitude, indicating both the interbedded nature of the formation and the depositional mechanism (down-slope turbidity currents or debris flows). The seismic relationship of the reflectors to volcanic massifs, together with the sandy mass flow and turbiditic nature of the sediments (wireline facies Bff2 and Sf3), attest to catastrophic deposition down-slope of the massifs. Away from the main volcanic centres, the internal seismic character of the Mohakatino Formation becomes more homogeneous (e.g., Wainui-1, Ariki-1), although the bottom (bounding) reflector of the formation is usually bright.

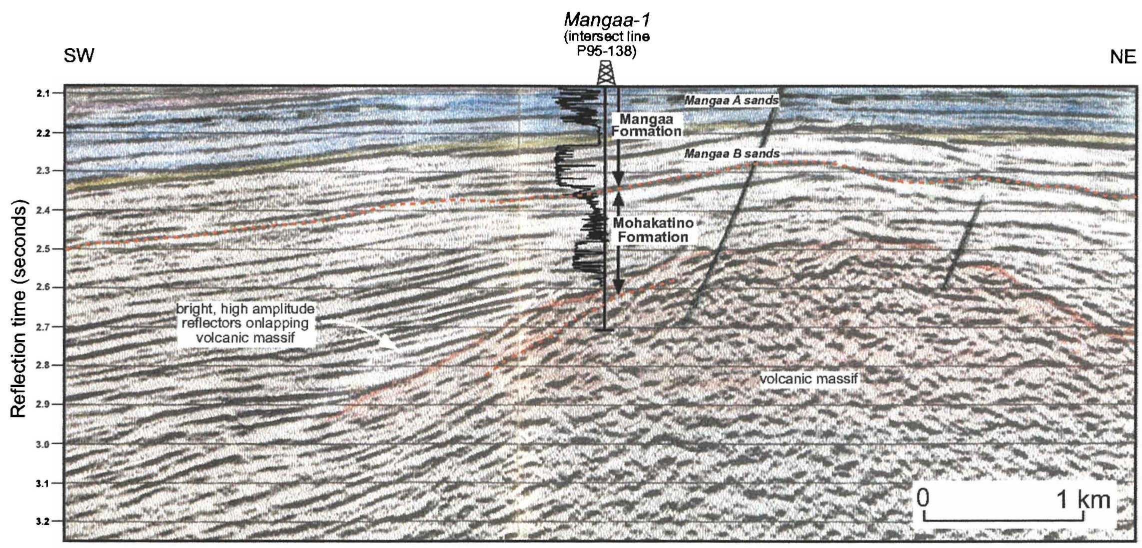


Fig. 5.4: An example of bright, high amplitude and continuous reflectors representing the Mohakatino Formation, onlapping Mangaa volcanc massif in the Northern Graben. Example is from seismic reflection profile P95-001, between shot points 940 (S) and 700 (N). Note the characteristic wireline (SP) response (see also Appendix 6).

5.4.2 Manganui Formation

Manganui Formation strata are contiguous in places with Giant Foresets Formation strata (e.g., Arawa-1, Taimana-1, Appendix 6). Within the study region, the Manganui Formation occupies an exclusively bathyal depositional environment. The seismic characteristic of the formation, reflecting its lithological homogeneity, is consistent with the predominantly muddy nature ascertained from well cuttings and wireline log signatures. Brighter internal reflectors within muddy Manganui Formation sediments often correlate with small sonic spikes and lower GR values on wireline logs, suggesting partial lithification and/or cementation (concretions?), or possibly thin condensed horizons. Most internal reflectors are of moderate to low amplitude, and many display lateral continuity, although more often than not reflectors are subparallel to wavy and discontinuous.

5.4.3 Ariki Formation

The Ariki Formation displays two distinct types of seismic reflection characteristics. At Tangaroa-1 and Te Kumi-1 the formation (represented by seismic unit A7) is bound by bright, high amplitude reflectors that correspond to marked deflections observed on GR and SP wireline logs, and lower sonic velocities (condensed wireline facies). This is a result of the velocity contrast between the marl and the underlying muddy Manganui/Mohakatino Formation. Between these boundaries is a reflection-free zone, a pattern characteristic of pelagic and hemipelagic sediments (e.g., Prather et al., 1998). However, in contrast the Ariki Formation at both Ariki-1 (seismic unit A4 and lower part of A8; top of unit A8 is time-equivalent to unit A7 at Tangaroa-1) and Wainui-1 (lower part of seismic unit A8), while displaying a bright basal reflector and similar wireline characteristics, occurs within a much more chaotic zone, characterised by wavy or hummocky, discontinuous and moderate amplitude reflectors, and an indistinct upper boundary (refer to Ariki-1 and Wainui-1, Appendix 6).

At both Ariki-1 and Wainui-1, the Ariki Formation incorporates slightly older sediments than at Tangaroa-1 or Te Kumi-1. Both of the former sites are basinward of a low-relief paleohigh, while the latter are located at or near the apex of paleo-highs. Given the chaotic nature of the upper part of the seismic reflection pattern, and the location of both Ariki-1 and Wainui-1 in relation to the paleohigh, as well as evidence of onlap, it is apparent that debris flows occurred off these subtle paleohighs during the Late Miocene-Early Pliocene and have contributed to the Ariki Formation.

While this discussion suggests that development of the Ariki Formation may be more complex than previously thought, involving hemipelagic and pelagic sedimentation and mass emplacement processes, the top of the seismic units representing the formation can be mapped between wells, and across the western part of the study area (Fig. 5.5), pinching out against or near faults that bound the western margin of the Northern Graben. The Ariki Formation *per se* cannot be traced with confidence eastward past these faults and into the Northern Graben – an indication that graben subsidence limited the distribution of the Ariki Formation due to more active siliciclastic sedimentation within the graben. A thin tuffaceous marly unit of equivalent age to the Ariki Formation occurs immediately above the Miocene-Pliocene unconformity at Kahawai-1, east of the graben (basal part of seismic unit A16). On seismic reflection profiles AR90-445-103 and AR89-446-108 (west of the graben) the Ariki Formation appears to abruptly end against a relatively thin, reflection-free unit bound by very bright, high amplitude reflectors, but due to the discordant nature of the upper boundary, particularly at Wainui-1 where the formation cannot be confidently tied into any other well, the southern extent of the formation is uncertain (refer to Fig. 5.5).

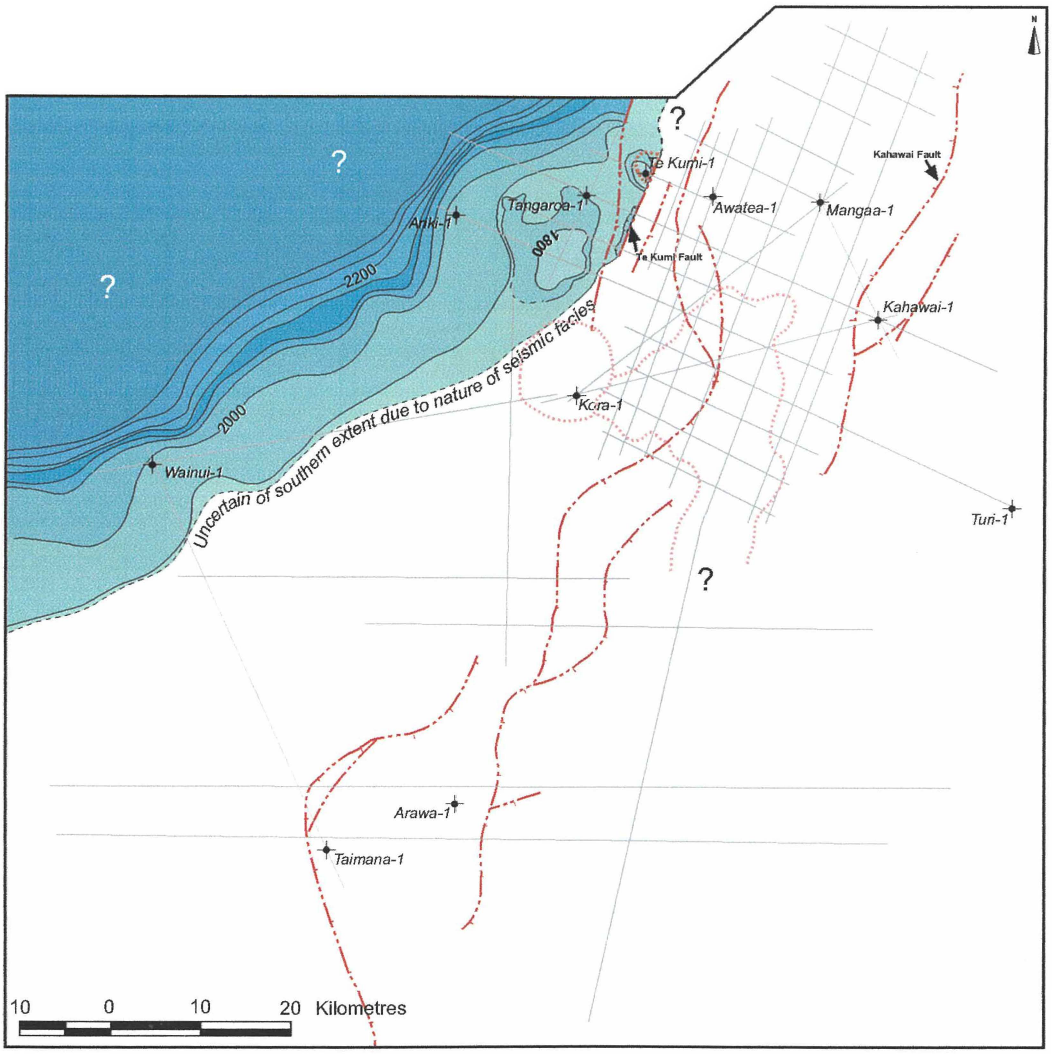


Fig. 5.5: Maximum extent of the Ariki Formation (structure-contour map in TWT, 50 msec intervals). To the north, the formation is abruptly truncated against Te Kumi Fault (downthrown to the east), while its more southern limits are harder to define, due to the chaotic and disrupted seismic reflection character on lines (e.g., Wainui-1, Appendix 6). Pink dashed lines indicate possible extent of submarine exposure of Miocene volcanic massifs at this time.

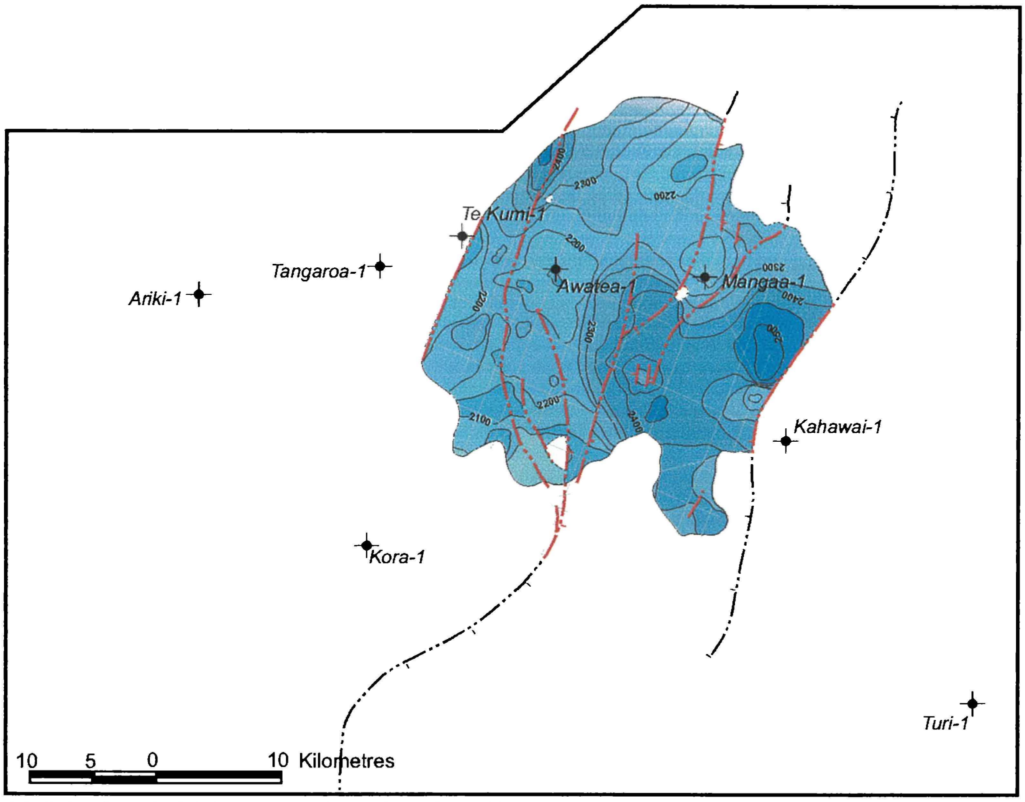
5.4.3(i) Other significant condensed units and implications for graben development

The Ariki Formation is the most easily distinguished of all lithologies by its wireline signature and lithology. The wireline motif indicative of condensed marly horizons, coupled with well report descriptions, has led to the identification of two other significant and possibly condensed marly horizons in the Awatea-1 and Mangaa-1 well sections. As discussed in Chapter 4, a marly to silty unit occurs immediately below the Miocene-Pliocene unconformity, and another (higher) unit separates lower and upper sandstone beds of the Mangaa Formation.

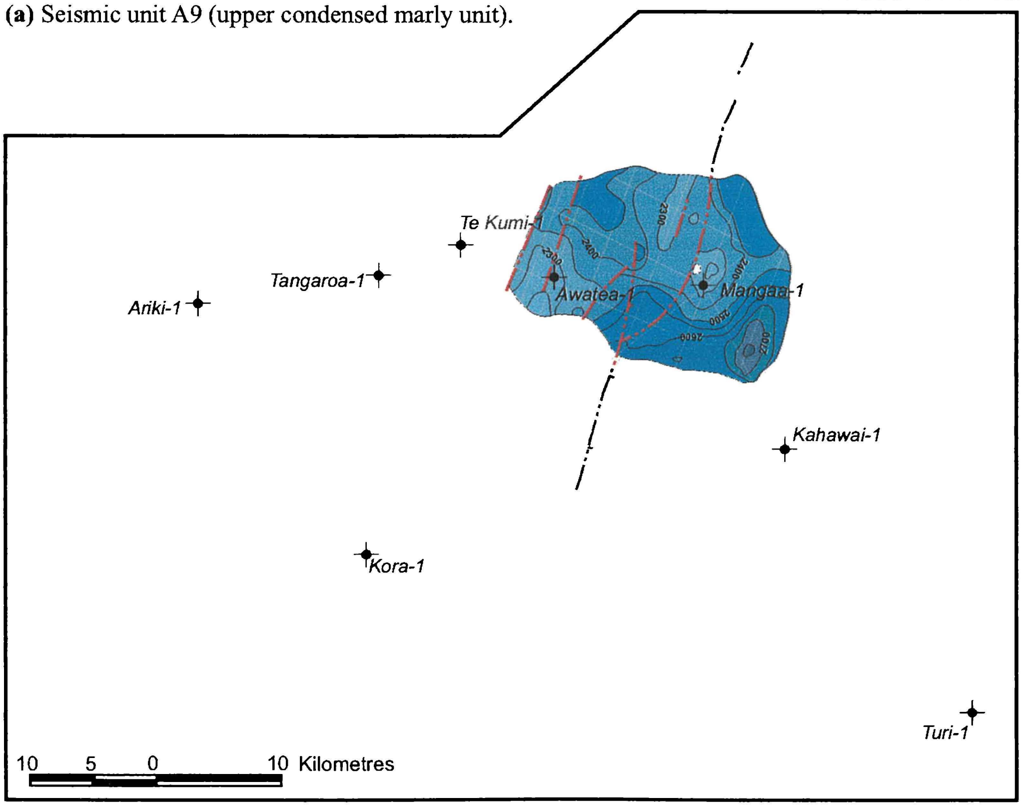
At Awatea-1, the upper unit (between 2499 and 2541 m) is very marly at the base, grading upward into siltstone. The marly part of the unit corresponds to seismic unit A9. At Mangaa-1 (between 2525 and 2604 m, also represented by seismic unit A9) this unit is described as a very calcareous marl (Hematite Petroleum (NZ) Ltd., 1970). On seismic reflection profiles, seismic unit A9 can be traced over much of the central part of the Northern Graben (Fig. 5.6a), although its seismic character changes with distance from the well sites. Close to Awatea-1, where it is best developed, it displays similar characteristics to the Ariki Formation observed at Tangaroa-1 and Te Kumi-1, with the lower bounding horizon being particularly distinctive (i.e., bold and continuous), with a quiet, reflection-free internal character. To the south in particular, the bounding reflectors become less bright, at times become discontinuous, and in places subparallel internal reflectors appear, indicating that the unit is becoming less homogeneous.

The lower marly unit, which underlies the Mangaa Formation, occurs below the lowest seismic unit mapped in the Northern Graben (seismic unit A9). As it is a potentially significant interval, occurring in conjunction with the Miocene-Pliocene unconformity (paraconformity), it has also been mapped (Fig. 5.6b). At Mangaa-1 the lower unit (between 2786 and 2822 m) is identified as a distinctly marly interval (on wireline log response and also identified in King and Thrasher, 1996, their Appendix 3). At Awatea-1 (~2695 – 2704 m) the interval is siltier, but can be directly correlated to the same interval at Mangaa-1. To the south it is apparently truncated by seismic unit A9, which drapes across the sandier basin fan floor lobes of the Mangaa Formation. In other directions, it conformably pinches out.

Seismic unit A9 is of comparable age to the Ariki Formation at Tangaroa-1 and Te Kumi-1, but cannot be directly correlated to this formation at these sites. This is possibly due to movement on faults bounding the western edge of the Northern Graben, or it may be that these condensed units are stratigraphically unrelated. The lower interval is of a similar age to the Ariki Formation at Ariki-1 and Wainui-1 (Late Tongaporutuan to Early Kapitean), but cannot be correlated far to the west.



(a) Seismic unit A9 (upper condensed marly unit).



(b) Unnamed seismic unit, base Mangaa Formation (lower silty-marly unit).

Fig. 5.6: Structure-contour maps (TWT, 50 msec intervals) showing the extent of the Ariki Formation age-equivalent marls. (a) seismic unit A9, between Mangaa B and Mangaa A sands; (b) unnamed marly unit at base of the Mangaa Formation. Both units are correlatable between Mangaa-1 and Awatea-1, but cannot be correlated to the Ariki Formation.

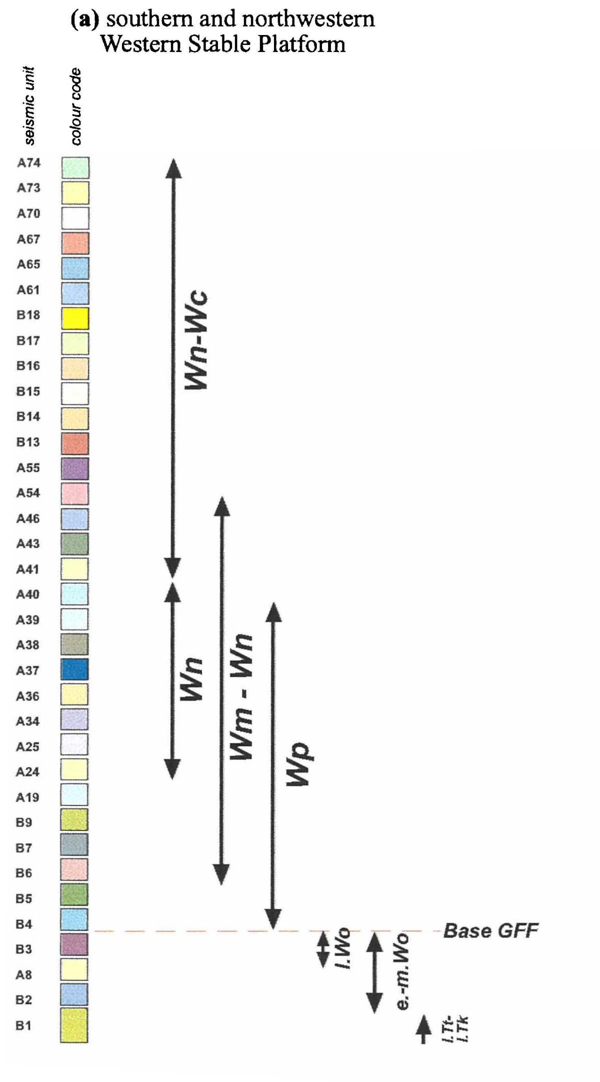
Both units may represent periods of fan abandonment and sediment starvation and, as with the Ariki Formation, are time-rich compared to the intervening sand-dominated intervals.

5.5 Chronology and broad depositional geometry of the Plio-Pleistocene succession

The ability to transfer biostratigraphic stages on to seismic sections is a crucial step in defining the chronological order of events that have contributed to the outbuilding of the continental shelf and slope making up the Giant Foresets Formation. The chronology of the Late Miocene to Pleistocene succession is discussed in depth in Chapter 3. Fig. 5.7 presents a summary of the results. Only those seismic units represented in well sections are presented. For a full list of all units and relative ages, refer to Appendix 5d. Note that the chronology presented in Fig. 5.7 includes uppermost Manganui Formation (Early Pliocene; Arawa-1 and Taimana-1) and the Mangaa Formation (Pliocene; Awatea-1 and Mangaa-1), as well as the Giant Foresets Formation.

Seismic reflectors are essentially isochronous depositional surfaces, and are therefore chronostratigraphically significant, separating older strata below from younger strata above (Vail, 1987). While there are exceptions to this model (e.g., Christie-Blick et al. (1987) in Cross and Lessenger, 1988), seismic reflections may be thought of as geologic time boundaries. It therefore follows that a seismic reflector traced from place to place should be approximately the same age at any point along that reflector. However, with reference to Fig. 5.7, it is obvious that there is much overlap between well sites in terms of the New Zealand Stages applied to the sedimentary succession at each well site. The reasons for this are many, including the diachronous nature of the foresets themselves and complications resulting from fault movement, but the primary explanation must be attributed to lack of confidence in the biostratigraphic dating itself. Successive stages therefore overlap by reference to seismic units.

While this overlap exists, and does result in some problems, there is good control between Plio-Pleistocene sediments and Miocene sediments. The position of the Miocene-Pliocene unconformity (paraconformity) is well constrained, and Miocene ages do not overlap Plio-Pleistocene ages, whether strata are conformable (i.e., Taimana-1, Arawa-1) or unconformable (e.g., Wainui-1, Ariki-1).



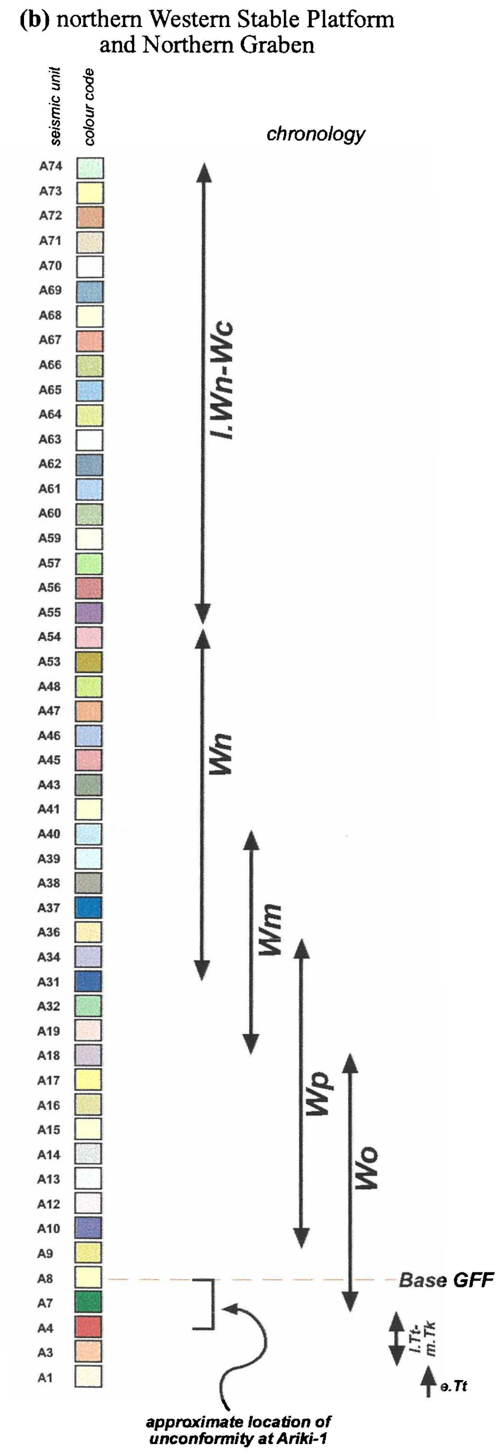


Fig. 5.7: Composite biostratigraphy (based on New Zealand Stages) for Western Stable Platform and Northern Graben seismic units. **(a)** northwestern and southern Western Stable Platform (Arawa-1, Taimana-1, Wainui-1); **(b)** northern Western Stable Platform and Northern Graben (Kora-1, Ariki-1, Tangaroa-1, Te Kumi-1, Awatea-1, Mangaa-1, Kahawai-1, and Turi-1). The overlap in stages (ages) reflects the uncertainty in the assignment of biostratigraphic stages to the seismic units, as discussed in Chapter 3.

5.5.1 Isopach maps

Although the strata corresponding to particular biostratigraphic stages cannot be uniquely defined, by choosing a mid point between seismic units the generation of isopach maps can be achieved. While these are not as accurate as one would like, they nevertheless illustrate the gross depositional patterns through the Plio-Pleistocene (Fig. 5.8; note that these maps include everything above the Miocene-Pliocene unconformity, including the Mangaa Formation, and to the south, conformable Opoitian Manganui Formation sediments). Figure 5.8 clearly shows that during the Opoitian, sedimentation was focused south of the study area (Manganui Formation), with some deposition occurring within the newly created graben (Moki C2/Mangaa B sands, part of Moki C1/Mangaa A sands). To the north, thick intervals of Waipipian and Mangapanian sediments attest to an advancing foreset front (bottomset component of the Giant Foresets Formation), although sedimentation was still focused in the southwestern part of the study area, with local deposition (Mangaa Formation) within the Northern Graben. A particularly thick package of sediment also occurs to the northeast of Arawa-1, in a depositional low created by concomitant movement along the Cape Egmont Fault Zone (Soenander, 1991).

The foreset front had advanced rapidly into the study area by the Early Nukumaruan, and the depositional loci changed from the southern part of the study area to the northern part, with progradation occurring along a roughly west-northeast lineament between Wainui-1 and Mangaa-1. This depositional front was widest across the Northern Graben, attesting to the influence the graben still had in attracting sediment. By the late Nukumaruan, the loci of deposition had continued to migrate north and westwards, with the depositional front centered to the northwest of the study area. Present-day active deposition is thought to be predominately north of Tangaroa-1 (Beggs, 1990).

5.5.2 Mangaa Formation

The Pliocene Mangaa Formation has been intersected in only two wells: Mangaa-1, after which the formation was named, and Awatea-1. Both of these wells are located in the central part of the Northern Graben, in which the thickest accumulation of post-Miocene sediment is recorded (~ 2700 m). While the Mangaa Formation is a separate unit to the Giant Foresets Formation, distinguished by a blocky and tabular wireline motif and coarse siliciclastic lithology, it essentially represents the first phase of aggressive sedimentary deposition within the northern part of Taranaki Basin.

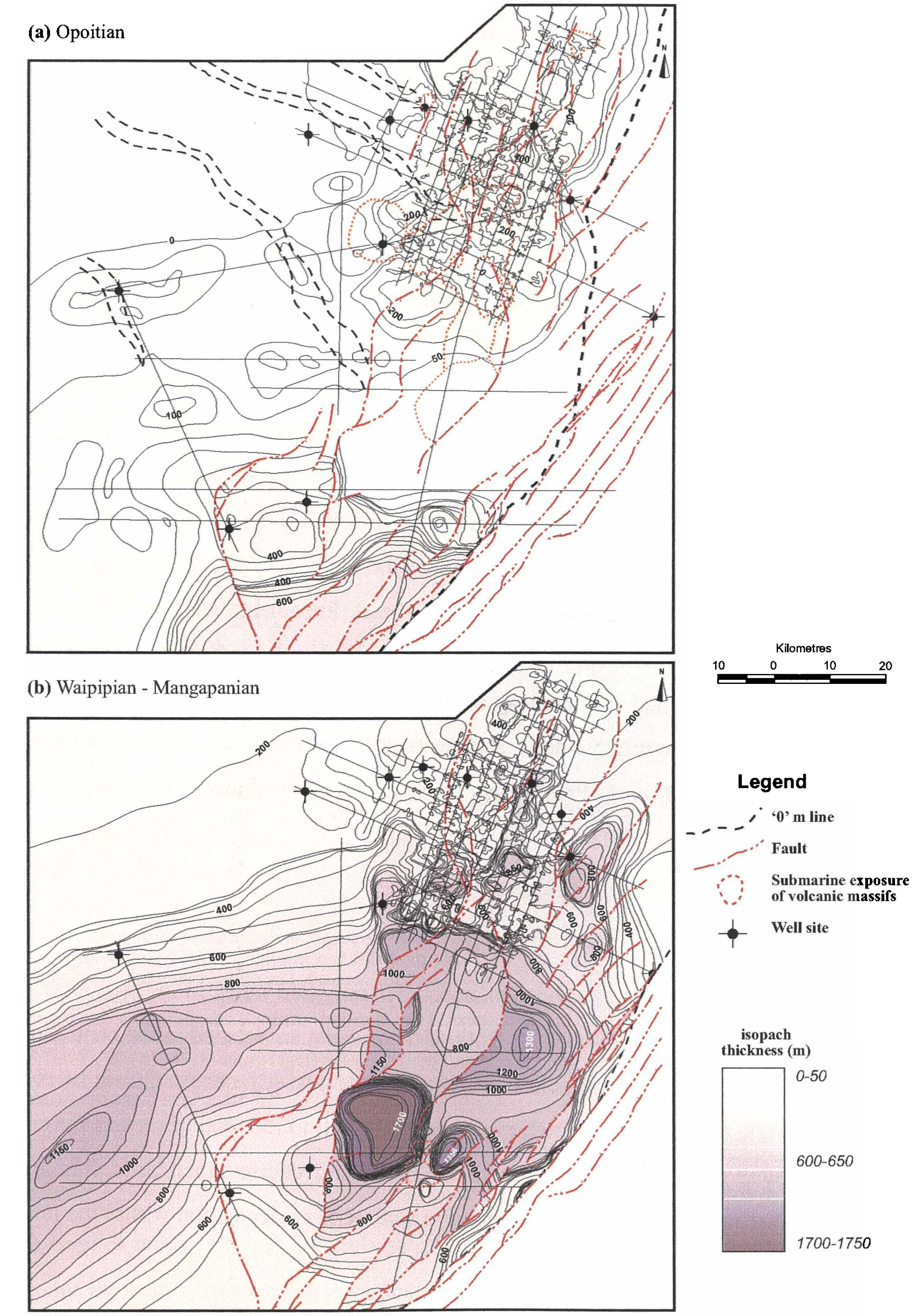


Fig. 5.8a-d: Isopachs showing thickness distribution and depositional patterns within the Plio-Pleistocene Giant Foresets Formation sedimentation for four intervals. Thickness in metres. '0' m line is point at which sediment is no longer present as a result of post-depositional erosion. Isopach contours in 50 m intervals. Top Miocene channels of Thrasher and Cahill (1990) shown on 5.8a.

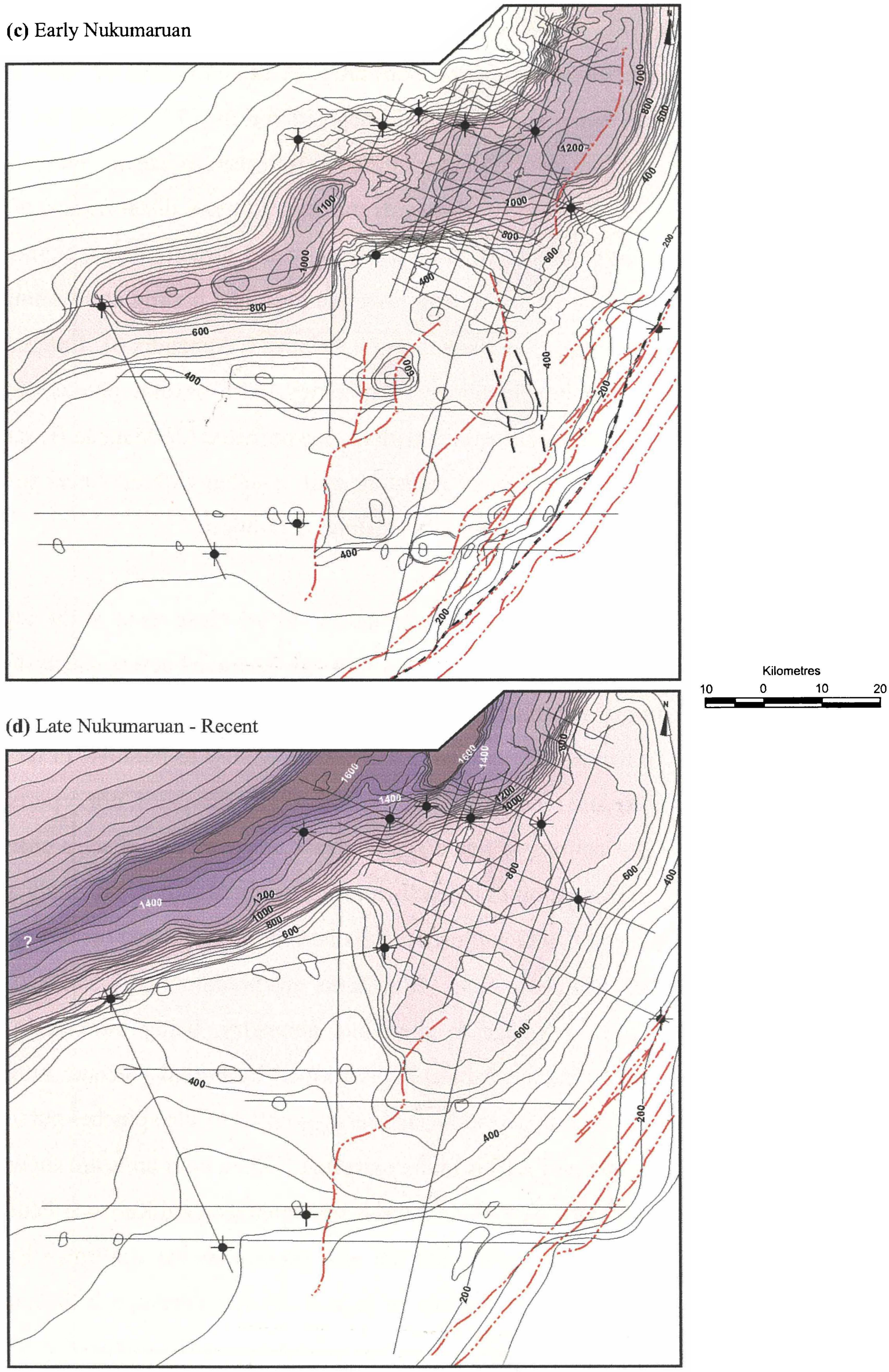


Fig. 5.8 continued.

Overall, the blocky nature of the Mangaa Formation, as displayed by wireline logs (SP and resistivity), is mirrored by the reflection character of seismic units A10 to A18. Each seismic unit is characterised by quiet (relatively reflection-free) intervals representative of mass flow units (wireline facies Bff1) deposited along the main depositional axis of a basin-floor fan complex, while the internal bright, parallel and continuous reflectors relate to intercalated mudstone beds (hemipelagic wireline facies). Towards the top and more distal fringes of the formation, reflectors become less parallel and more hummocky and discontinuous, emulating the less blocky nature observed on equivalent depth wireline logs, and hinting at a waning of the basin-floor fan depositional system (see Mangaa-1 and Awatea-1, Appendix 6). Externally, the Mangaa Formation displays a low mounded morphology, with several phases of fan development indicated, most noticeably associated with the lower (Moki C2/Mangaa B) sands. Seismic unit A9, which drapes across the lower sandstone unit, together with the lower part of seismic unit A10, separate the two major pulses of sandstone deposition.

The Mangaa Formation-Giant Foresets Formation boundary occurs close to or at the top of seismic unit A18 in the Northern Graben. While this unit can be traced across much of the northern part of the P95 seismic grid, and is moderately sandy at both Awatea-1 and Mangaa-1, at other well sites (e.g., Te Kumi-1, Tangaroa-1) it is distinctly silty to muddy. In fact, as discussed in Chapter 4, no other site records the same wireline characteristics, nor reports the same or similar lithological distinctions as noted at Mangaa-1 or Awatea-1. Facies changes limit the extent of the Mangaa Formation to the graben in the northern part of the study area. This is highlighted in Fig. 5.9a and b, which illustrate the distribution and thickness of seismic units A10 and A15 of the Mangaa Formation. Both units are predominantly sandy (from wireline interpretation) and form a wedge to sheet-like geometry, being thickest in the structurally lower regions, and thinner over paleohighs formed by buried volcanic massifs. Both seismic units pinch out to the north and south, and seismic unit A15 also pinches out to the west. The western limit of seismic unit A10 is fault controlled. While both units are shown as being truncated at the large fault to the west of Kahawai-1 (informally called Kahawai Fault for ease of reference), it is probable that these units are also present on the upthrown block, pinching out before Kahawai-1. However, because of lack of seismic control, it is difficult to confidently correlate across the fault at depth.

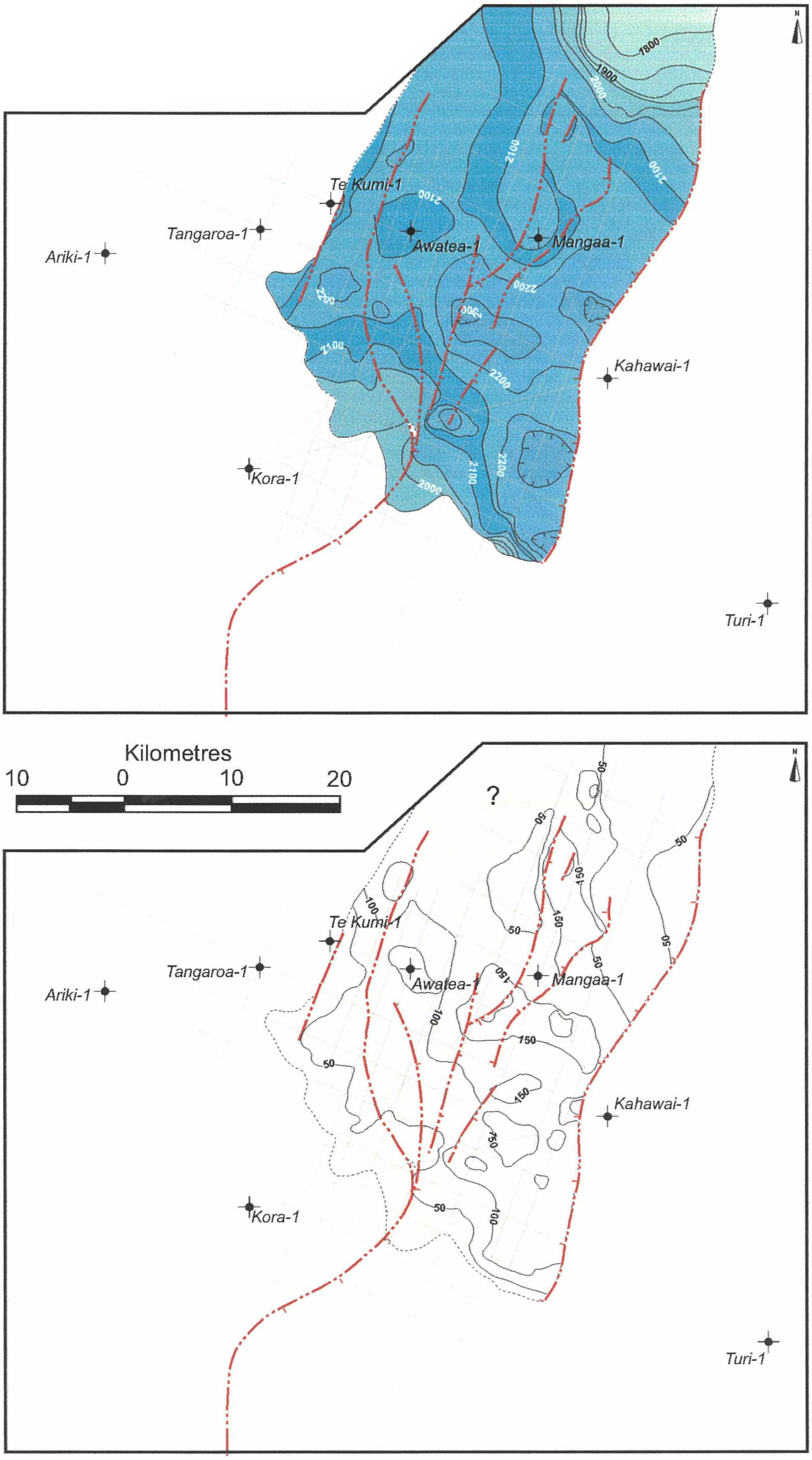


Fig. 5.9a: Structure-contour map (top; TWT, 50 msec intervals) and isopach map (bottom; 50 msec intervals) of seismic unit A10, Mangaa Formation, Northern Graben.

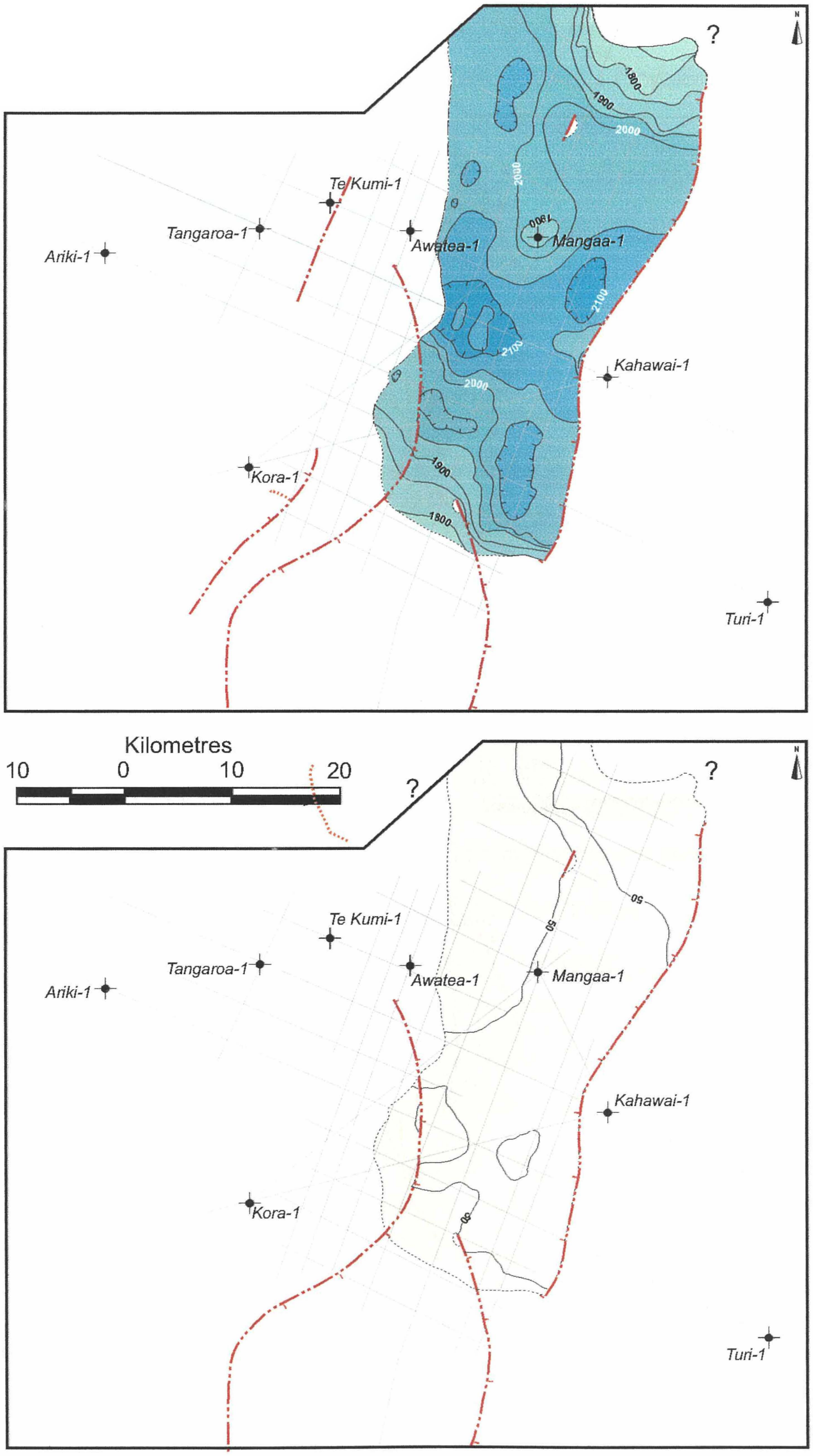


Fig. 5.9b: Structure-contour map (top) and isopach map (bottom) of seismic unit A15, Mangaa Formation, Northern Graben.

5.5.3 Seismic reflection characteristics of the Giant Foresets Formation

5.5.3(i) Definition of the base of the Giant Foresets Formation

The base of the Giant Foresets Formation is defined from well log character, often associated with a generally homogeneous motif that displays an overall upward-coarsening trend. This study follows the position of the base of the Giant Foresets Formation (m bKB) as defined on well sections from composite well reports, and King and Thrasher (1996). In New Zealand petroleum industry literature, it is typically associated with a bright high-amplitude and continuous horizon identified on seismic reflection profiles. This flat to gently dipping reflector horizon is often termed the 'yellow' horizon (e.g., Thrasher and Cahill, 1990) after the industry-standard colour used to delineate it. The yellow reflector horizon often separates underlying Miocene strata from overlying Plio-Pleistocene strata, and is therefore presumed to be synchronous with the regional Miocene-Pliocene unconformity.

In the study area, the seismic base of the Giant Foresets Formation occurs at or near the top of seismic unit A18 within the Northern Graben (refer to Enclosure 3), and near the top of seismic unit B3 on the Western Stable Platform (Fig 5.10) or the next lowest horizon where these seismic units pinch out. In the vicinity of Ariki-1 and Wainui-1, the base of the Giant Foresets Formation equates to the yellow seismic horizon. However, problems arise in assuming that the yellow horizon is always associated with the base of the Giant Foresets Formation. This occurs because part of the Plio-Pleistocene succession in fact comprises uppermost Manganui Formation strata in the vicinity of Arawa-1 and Taimana-1, and Mangaa Formation in the Northern Graben. The difference in down-hole depth between the seismic position of the yellow reflector horizon, and the actual depth to the base of the Giant Foresets Formation at each well site is presented in Table 5.1.

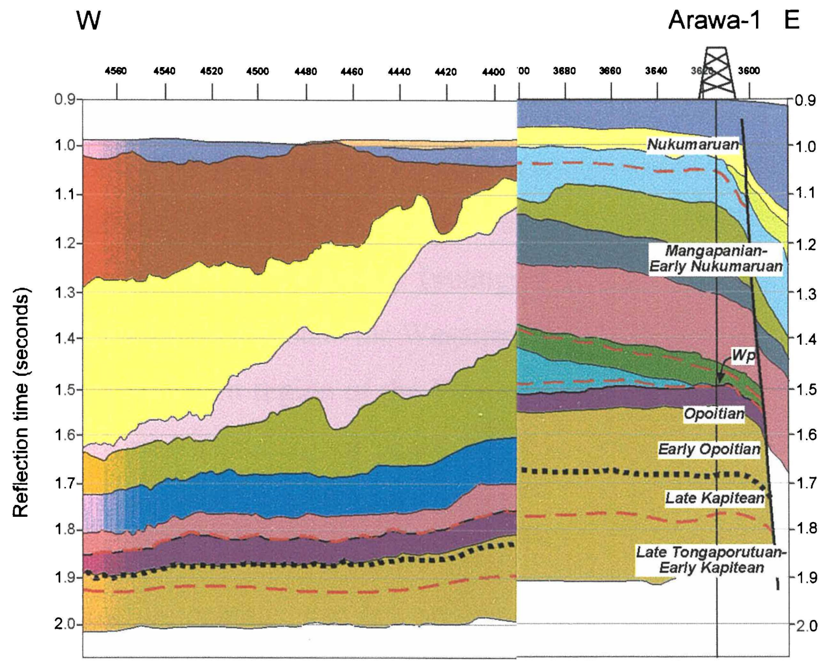
Across the northwestern parts of the study area (e.g., in the vicinity of Ariki-1 and Wainui-1), the yellow reflector is a useful guide in determining the approximate depth of the base of the Giant Foresets Formation. In all other areas, care must be exercised when delineating the base of the Giant Foresets Formation in the absence of well sections. For example, the base of the Giant Foresets Formation at Arawa-1 and Taimana-1 occurs within Opoitian to Waipipian-aged strata, and coincides with the top of seismic unit B3, which on seismic reflection profile NM 17 and AR90-445-103 is associated with a bold reflector horizon. As illustrated in Fig. 5.10, the top of seismic unit B3 is 150-200 msecs above the actual position of the yellow seismic reflector horizon at Arawa-1 and Taimana-1, and occurs within conformable uppermost Manganui Formation strata. Further to the northwest, toward Wainui-1, the yellow reflector horizon merges with the top of seismic unit B3, then defines the top of the Ariki Formation (within seismic unit A8). At all sites, the yellow reflector horizon separates Miocene from Pliocene strata. In the Northern Graben, the base of the Giant Foresets Formation is Waipipian in age (e.g., Awatea-1, Mangaa-1), and the yellow seismic reflector horizon occurs at the base of the Mangaa Formation, again separating Miocene and Pliocene strata.

Table 5.1: Depth to 'yellow' reflector horizon (data supplied by Geosphere Exploration Ltd., Lower Hutt) versus depth of the lower Giant Foresets Formation boundary (m bKB).

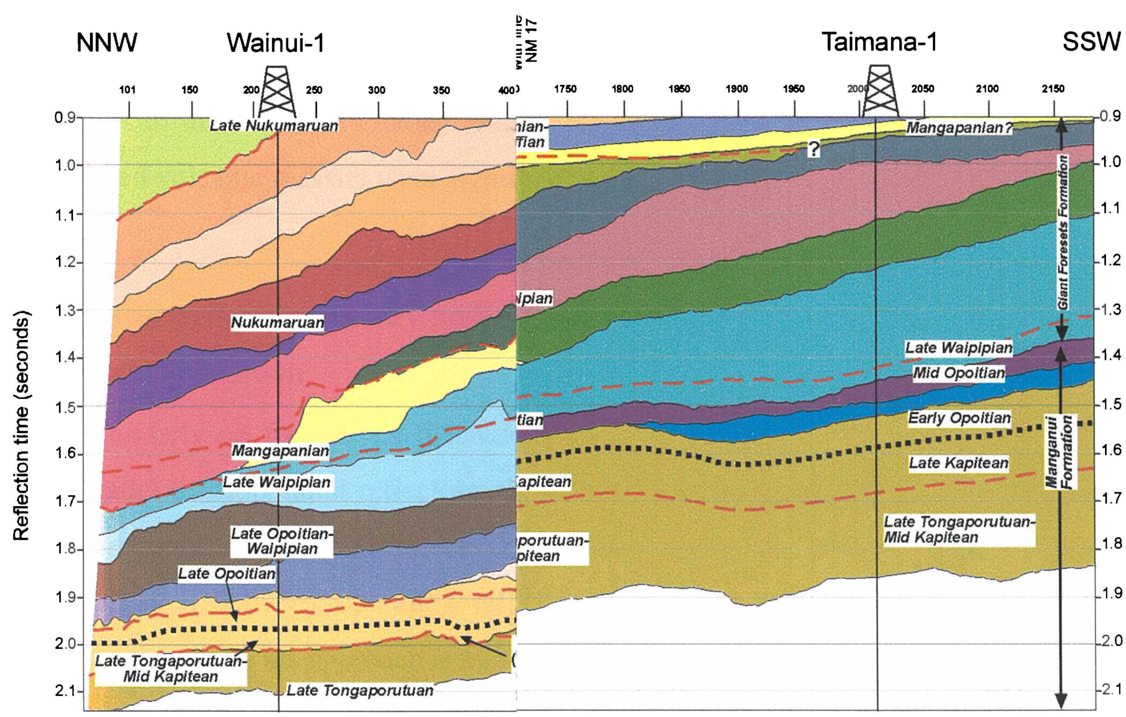
Well	Depth to yellow reflector (m bKB)	Depth to base Giant ForesetsFormation (m bKB)	Basal seismic unit of Giant Foresets Formation
Arawa-1	1795	1541	top B3
Ariki-1	2162	2147	A8
Awatea-1	2327	2107	top A18
Kahawai-1	1931	1931	top A16
Kora-1	1781	1781	A17
Mangaa-1	1990	2605	base A19
Taimana-1	1700	1496	top B3
Tangaroa-1	2085.3	2025	top A7
Te Kumi-1	2079	2000	top A7
Turi-1	NP	259	base A61
Wainui-1	2215	2215	A8

NP – not present

Correlation of seismic reflectors between the Western Stable Platform and the Northern Graben is difficult because of numerous volcanic edifices aligned along the axis of the Northern Graben, over which lies the Plio-Pleistocene succession. East of the graben, sediments are truncated by a young erosion surface (Manganui Platform), as the succession is progressively stepped-up over the Turi Fault Zone.



(a) NM17

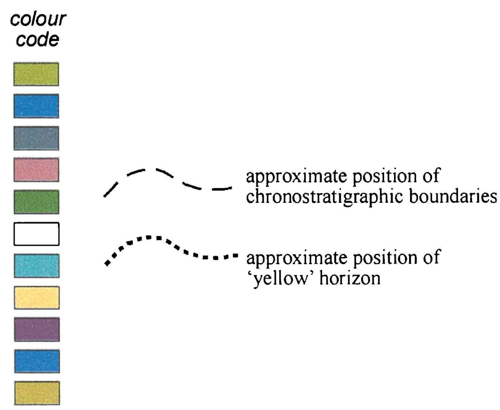


(b) AR90-445-103

;mic
iit
0

Fig. 5.10: Interpreted chronostratigraphy betwee,
reflector horizon marks the approximate Miocen
Formation (which on seismic lines AR90-445-103
display relatively good chronostratigraphic correl
nature of the foresets themselves, reflecting the v
of an inability to accurately constrain biostratigrap
at Arawa-1. Note the mounded nature of seismic

Vertical scale exaggerated 4x. See Fig. 5.2 for lo



Because of the somewhat diachronous nature of the base of the Giant Foresets Formation from well site to well site (as described above), creation of a structure contour map (to base Giant Foresets Formation) was more complex than simply mapping a particular horizon. Instead, the resultant map (Fig. 5.11) has of necessity crossed several seismic reflectors in the Northern Graben region to force a correlation between wells, with the basal age of the formation changing from the central part of the graben (youngest at Awatea-1 and Mangaa-1) towards the outer margins (Kahawai-1) and on to the Western Stable Platform (Ariki-1, Tangaroa-1, Te Kumi-1). Over the northwestern region of the Western Stable Platform, across the volcanic-dominated axis of the Northern Graben, and across the Turi Fault Zone, the base Giant Foresets Formation map is consistent with the yellow reflector horizon as mapped by Thrasher and Cahill (1990).

While a structure contour map is a present-day depth to a horizon, the map presented in Fig. 5.11 incorporates elements of the broad topography (submarine structure) of the study area that developed prior to or early on during the main phase of foresets progradation. It shows that, prior to deposition of the Giant Foresets Formation, the deepest areas in northern Taranaki were along the axis of the Northern Graben, while several topographic highs created by buried or partially buried volcanic massifs were significant features.

5.5.3(ii) Characteristics, geometry and distribution patterns of seismic units within the Giant Foresets Formation

Unlike the underlying formations discussed earlier, which are generally characterised by smooth, parallel to subparallel reflectors, the Giant Foresets Formation is characterised by a diversity of external and internal reflector configurations. It is this diversity of configuration patterns, packaged between bounding reflectors, which are used to differentiate between discrete seismic units. This approach is more of a stratigraphic one, and differs from the normal seismic interpretation approach of defining sequence boundaries and systems tracts (*sensu stricto* Vail, 1987) for several reasons:

1. while the Giant Foresets Formation displays remarkably distinct prograding clinoforms on seismic reflection profiles, sequence-defining recognition criteria (reflector terminations) are often difficult to recognise;

2. lack of well control, lack of sampling through topsets, and the thickness of topsets relative to foresets means that only the limited parts of the Giant Foresets Formation that include depositional slope succession can be used to delineate sequences.

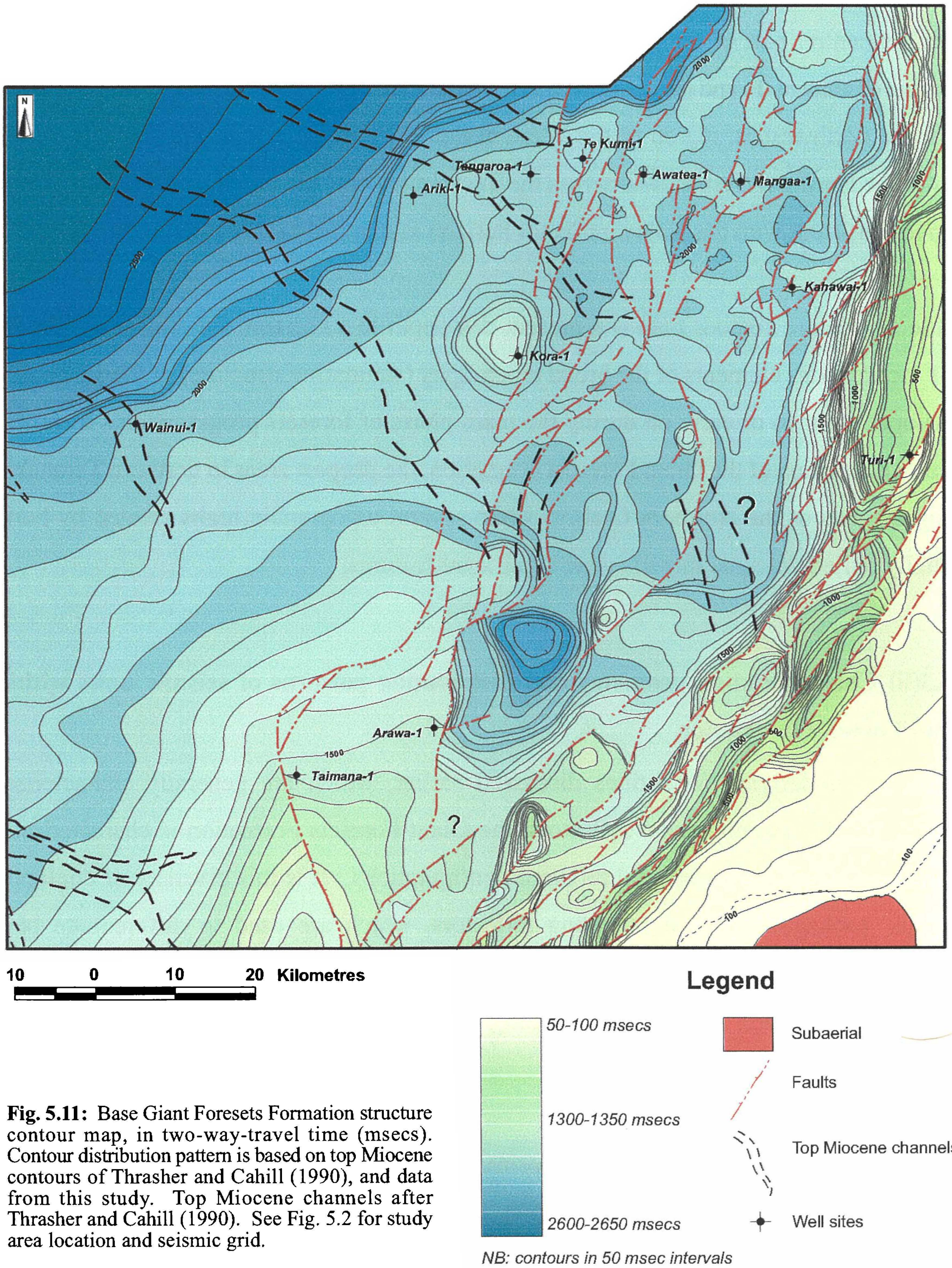


Fig. 5.11: Base Giant Foresets Formation structure contour map, in two-way-travel time (msecs). Contour distribution pattern is based on top Miocene contours of Thrasher and Cahill (1990), and data from this study. Top Miocene channels after Thrasher and Cahill (1990). See Fig. 5.2 for study area location and seismic grid.

Subsequently, the following is a description of individual seismic units or packages of seismic units as they appear on seismic reflection profiles within the study area. Section 5.6 examines in more detail the internal characteristics in terms of their depositional environment and mode of deposition, which are later related to their position on the relative sea level curve (5.8).

Seismic units of the Northern Graben and eastern part of the Western Stable Platform

Seismic units of the Giant Foresets Formation vary markedly from bottom to top and, based on the few wells that intersect various units, tend to echo the interpreted lithology. Seismic unit A18 is the first unit of the Mangaa Formation that is also included in the Giant Foresets Formation at other sites. This unit is thickest in the central part of the Northern Graben, and thins in all directions. It is also the first unit in the post-Miocene succession that does not appear to have been severely affected by faulting. Internal reflectors are reasonably continuous, and as with internal reflectors of the underlying sandy intervals within the Mangaa Formation, are of relatively high to moderate amplitude. This seismic unit conformably pinches out in most directions (does not occur much further west than Tangaroa-1), although to the south, because of distortion over volcanic massifs, it is unclear how far this unit extends.

Seismic unit A19 is bound by bright, continuous and moderate to high amplitude reflectors, and displays a variety of internal reflector configurations. These range from numerous parallel to subparallel, continuous and moderate to high amplitude reflectors in the central part of the Northern Graben, to hummocky and disorganised reflectors towards the outer margins of the graben and on to the Western Stable Platform. This more chaotic reflector pattern is most noticeable on seismic reflection profiles P95-158 (Enclosure 3) and P95-168, where the upper bounding reflector becomes more discontinuous, and the internal reflectors become highly disrupted. The change in both bounding and internal characteristics of this unit (and also of the underlying seismic unit A8, and overlying seismic units A31, A42, A44) are attributed to slump or debris flow deposits accumulating downslope of the paleo-high evidenced on seismic reflection profile P95-158.

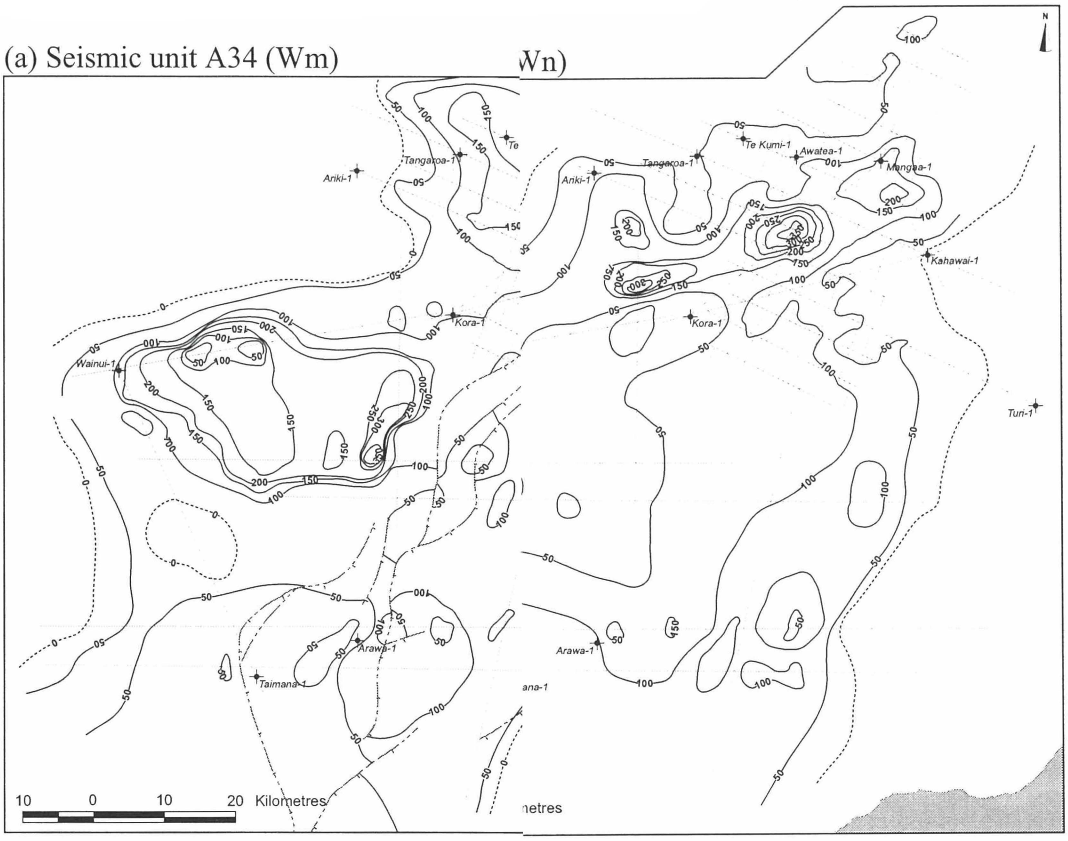
The first units to display foresetting in the Northern Graben are seismic units A34 to A40. Most of these units display a bottomset (basin facies), foreset (slope facies), and topset (shelf facies) component. Several of these units, particularly seismic units A34, A39, and A40 (Fig. 5.12a) are extensive, and form a broad foreset front from Wainui-1 in the southwest to Kahawai-1 in the northeast. The thickest part of the foreset front is on the Western Stable Platform, although by the Nukumaruan (seismic unit A40) lobes of sediment had advanced into the northern part of the study area. Seismic units A36 to A38 are much less extensive, with the thickest foreset part also restricted to the more western parts of the study area. This grouping of seismic units is characterised by bounding reflectors that are often discontinuous, and vary markedly from bright, high amplitude reflectors to discontinuous, disrupted and low amplitude reflectors. Occasionally, lower boundaries display shallow scour features, indicative of rotational slumping (see section 5.6.3(iii)). Internally, these units are characterised by discordant and disrupted to chaotic reflector configurations, with hummocky to mounded profiles and a distinct lack of organization.

Seismic units A41 and A43 display similar internal seismic attributes to the previous units discussed, but tend to be bound by more continuous, higher amplitude reflectors. As with the previous units, they are laterally extensive (e.g., Fig 5.12c, 5.12b), although the foreset part of the unit is often not well developed due to incision by overlying units. Both units, particularly seismic unit A41 exhibit external mounded geometries at the toe of the foreset wedge (see Enclosure 3), and numerous smaller internal mound forms with downlapping reflectors are observed.

A broadly channelised zone occurs between seismic reflection profiles P95-131 and P95-001 (refer to section 5.6.3(iv)) and has resulted in significant loss of the foresetted part of seismic units A46 to A53). They therefore tend to be thinner overall, and have less extensive bottomset components, although the topset components are still quite extensive. The topset (or shelfal) component of seismic unit A46 is in fact not only extensive (it is present over much of the study area) but is also very thick compared to most other units (often greater than 100 msecs), and includes numerous moderate amplitude parallel to subparallel reflectors indicative of an interbedded sequence. Internal reflectors within the foresetted part of these units are often low to moderate amplitude, discontinuous and wavy to hummocky. Some downlap on to hummocky or low-relief mounds is observed, although this is often difficult to pick because of the nature of the reflectors. Bounding reflectors are frequently low amplitude basinward of the shelf break, but tend to be brighter and higher amplitude landward of the shelf break. Other than seismic unit A46 these units tend to be restricted to the northern part of the study area (e.g., Fig. 5.13c).

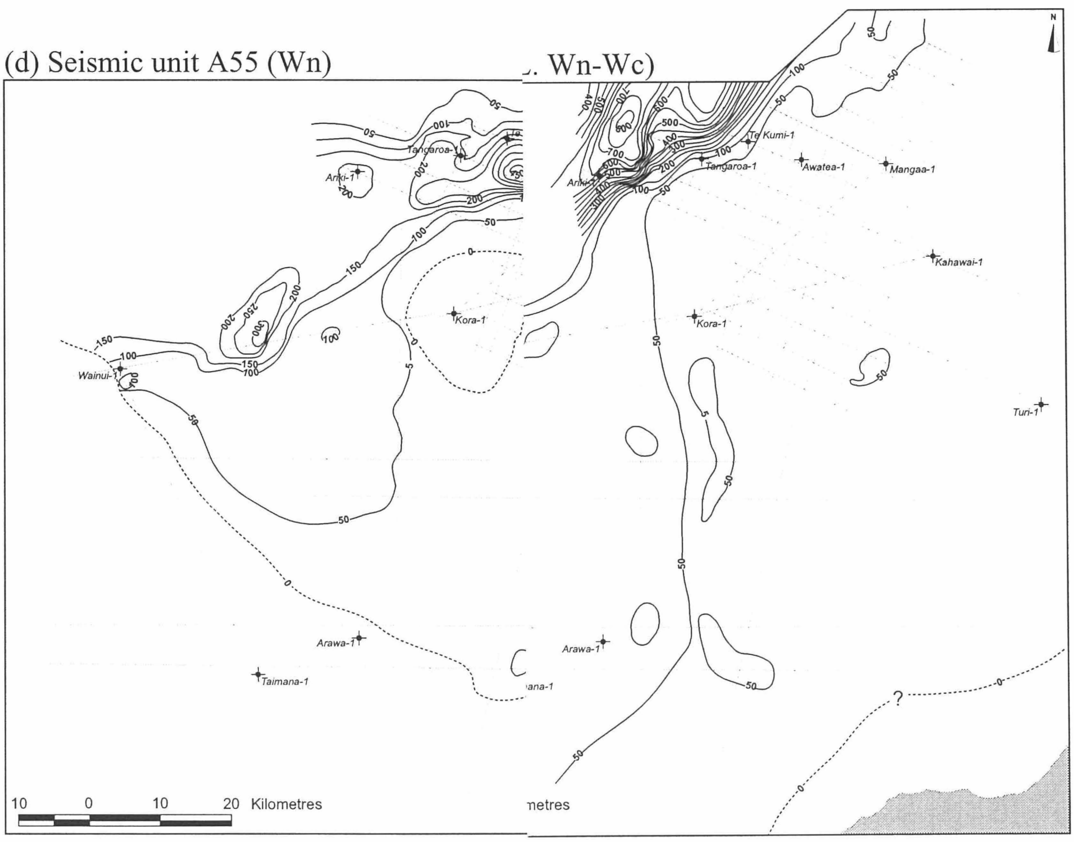
(a) Seismic unit A34 (Wm)

Wn)



(d) Seismic unit A55 (Wn)

. Wn-Wc)



Legend

Fig. 5.12: Isopach maps of selected seismic units wl e past which unit can no longer be mapped ....
from elongate and focused, to forming a more lineaı

Channels

re in 50 msec intervals

A distinct change in seismic character occurs between the seismic units discussed previously, and seismic unit A54. This unit has a much thicker foreset wedge, is bound by bright, continuous and high to moderate amplitude reflectors, and can be traced over much of the study area (Fig. 5.13d). What makes it particularly distinctive, however, is the conspicuously hummocky external profile displayed on dip lines particularly on P95-158 (Enclosure 3). This profile is indicative of a moderately large channel levee system (Emery and Myers, 1996), and the geomorphology may in fact be related to a wide channel that incises deeply into this and underlying seismic units. Roll-over of parallel internal reflectors towards the channel margin, and divergence of these reflectors away from the margin, would appear to substantiate this. The upper mounded boundary is further demarcated by the presence of a second strong reflector 20-30 msecs below it, separated by a prominent reflection-free zone. The presence of this 'doublet' (e.g., Weimer, 1990) and the consistent thickness of the unit (unaffected by any abrupt change in slope) are indicative of hemipelagic drape (Prather et al., 1998).

Seismic units A55 to A60 are characterised by relatively continuous, moderate amplitude external reflectors, while internally, reflectors display a number of attributes including lateral continuity, occasional internal downlap on to low-relief mounded profiles, and external mounded geometries at the base of slope. Foresets are characterised by occasional toplap, but more often than not, by concordant roll-over into the shelfal part of the unit. Shelf (topset) components are initially very thick (up to 200 msecs in the case of seismic unit A60) but rapidly thin landward of the shelf break. While some of these units are extensive, isopach maps illustrate that the main loci of deposition was quite restricted (Fig. 12d,e). This is particularly true for seismic unit A59, which forms a distinctly elongate lobate body, pinching-out rapidly towards the margins of the lobe. Coupled with extensive toplap (the most prominent for any unit), this unit is interpreted as a slope-delta fan.

Seismic units A61 to A70 are erosionally truncated by seismic units A71 to A73 before their contemporary shelf break, and are thus only represented by topset facies. Many of these units, most obviously seismic unit A66, thicken through depressions and thin over slight rises between P95-001 and P95-103 to the east and west respectively, and P95-138, P95-174 to the north and south. This feature is analogous to compensation style bedding. Many of the units, particularly the thicker ones such as seismic units A66, A67, A68, and A70 display a shingled, offlapping internal reflector configuration that is indicative of prograding sediment bodies on the continental shelf (Enclosure 3; Miall, 1997). Towards the south and east, these units either pinch out (e.g., seismic unit A66), or maintain a constant presence over much of the study area (e.g., seismic unit A67).

Distinctly different from the units they truncate, seismic units A71 to A73 are characterised by discontinuous, low to moderate amplitude bounding reflectors, deep incision into underlying units, and disrupted to highly chaotic internal reflectors. These units (assumed for the erosionally truncated unit A71) are very thick (e.g., Fig. 5.12f), and are probably a composite of more than one unit. However, subdivision is not possible because of the lack of defining reflectors. Landward of the shelf break these units rapidly thin, but maintain a constant presence over much of the northern part of the study area, and in the case of seismic unit A73, the entire study area (Fig. 5.13h).

Seismic unit A74 is the uppermost unit mapped, and is characterised by continuous, moderate to high amplitude bounding reflectors and numerous moderate to low, continuous and parallel to subparallel internal reflectors. This unit occurs immediately below a pair of bold reflectors that constrain the most recent phase of seafloor sedimentation.

Seismic units of the Western Stable Platform and southeastern part of the study area

Many of the seismic units discussed above have been mapped across the entire study area, but a number of seismic units are particular to the western and southern part of the study area. These units have been delineated by the prefix 'B'. Seismic units B1 to B3 belong to the uppermost part of the Manganui Formation; all other seismic units are part of the Giant Foresets Formation. In the vicinity of Arawa-1 and Taimana-1, several seismic units within the Giant Foresets Formation (B4–B7) display a broadly mounded external profile that is not observed anywhere else in the study area (e.g., Fig. 5.10). The lower boundary of seismic unit B4 is delineated by a bold, high amplitude reflector which serves to delineate the seismic base of the Giant Foresets Formation. Parallel to subparallel internal reflectors within seismic units B4-B7 downlap this basal reflector in both dip and strike lines (refer to Fig. 5.16d). These units have a restricted aerial extent (e.g., Fig. 5.12a; see also Soenander, 1991, 1992).

Seismic units above unit B7 display external geometries more characteristic of prograding clinoforms. Seismic units B8-B12 demonstrate a progressive steepening of the slope profile towards the north and west (Fig. 5.10). External profiles become less smooth, and more hummocky. Related to this steepening is a change in internal reflector characteristics, from coherent, continuous and moderately high amplitude reflectors, to more chaotic, disrupted and discontinuous internal reflectors. This suggests a slope profile that progressively steepens beyond the angle of repose, until finally slope failure occurs. These seismic units make up a broad band of units (including A38-A40) that display internal disruption and evidence of slump scours, which form a southwest-northeast swath across the central part of the study area.

On strike lines (HF 510, HF 11X, AR89-446-108), seismic units A34 to A40 display a distinctly hummocky external profile, with relief of up to 200 msecs. This lobate shape in transverse section gives the impression of fan mounds (e.g., Mitchum, 1985), however, the high relief and external geometry in dip section indicate that these are in fact foresets that have been locally eroded through channel incision.

Seismic units B13 to B17 are only mapped in the far western part of the grid (e.g., Fig. 5.13e). These pinch-out rapidly landward of the shelf break, and are characterised by hummocky or wavy internal reflectors in the upper to mid slope region and more gently rounded mounds, characterised by downlapping reflectors, towards the toe of the slope (see Wainui-1, Appendix 6). While many of the internal reflectors are discontinuous, some internal reflectors display well-developed lateral continuity. Bounding reflectors are bright, but often discontinuous. These units are generally thinner than the preceding units, and display lower slope angles.

The uppermost seismic units mapped in the western and southern part of the study area are seismic units A65, A67, A70, A73 and A74. Some of these units (e.g., A65 and A74) are only present across the Northern Graben as relatively thin units, but thicken considerably to the southwest of Ariki-1. Conversely, seismic unit A73 (Fig. 5.13f) is present only as a relatively thin unit compared to the thickness of this unit in the vicinity of Ariki-1.

5.5.3(iii) Seismic distribution pattern and correlative units

The Giant Foresets Formation is an extensive formation, and has been mapped on the basis of its seismic character from south of Taranaki Peninsula, to north of the study area (e.g., Stagpoole, 1997), and as far west as Tane-1. In the southernmost and easternmost regions, the formation as defined in well reports is relatively thin, in the order of a few hundred metres or less where recorded (e.g., Maui-2, Maui-4, Turi-1). As shown in Enclosure 1, the formation is unconformably cutout across the Taranaki Peninsula, but is inferred to be the lateral equivalent of the Whenuakura Subgroup (shelf deposits) and Tangahoe Mudstone (slope deposits) identified in many southern Taranaki and Wanganui Basin wells (Kamp et al., 2002). In other wells, lateral equivalents are simply referred to as undifferentiated sediments. The diachronous nature of the formation means that in parts, the basal Giant Foresets Formation is coeval with Manganui Formation sediments.

5.6 Major seismic reflection facies (architectural elements) within the Giant Foresets Formation

Seismic facies analysis involves the delineation and interpretation of various reflector attributes, including external geometry, continuity, amplitude, frequency, and interval velocity (see Appendix 5b and Enclosure 3; Sangree and Widmier, 1977; Cant, 1992). Recent studies discuss these facies in terms of architectural elements (e.g., Galloway, 1998; Stow and Mayall, 2000, their Fig. 7, pg 131), recognising several fundamental building blocks of deep basins, including hiatuses, mass-wasting features and their products, fan mounds, channel incision and tectonic features. Each of these elements may occur in a range of scales and within a hierarchy of similar features (Stow and Mayall, 2000). Subsequently, the seismic units discussed in section 5.5 are interpreted here on the basis of their external and internal reflector configurations, which are used to delineate seismic facies (architectural elements) and are further interpreted in terms of the environment of deposition in which they form, and the associated depositional mechanisms. Deposition relative to position on a sea level curve is alluded to here, and discussed more fully in section 5.8.

5.6.1 Efficient versus inefficient systems: the importance of sediment supply and lithology

The type of sediment within a deepwater basin system (i.e., mud-rich systems versus sand-rich systems) has a profound effect on slope morphology, channel systems and sediment distribution, and type of transportational mechanism. Table 5.2 lists the various sedimentalogical characteristics of deep marine clastic systems, discussed in Reading and Richards (1994) and Emery and Myers (1996). As the following sections will demonstrate, the Giant Foresets Formation evolved as a mud- and mixed mud/sand-rich depositional system.

Table 5.2: Sedimentological characteris

	Single point-source submarine categorised by dominant grain ain size			
	Mud (MF)*	Mud/sand ;and (MSF))	Sand (SA)	Gravel (GA)
Size	Large	Large-modrate	Small	Very small
Slope-gradient (m km-1)	Low 0.20 – 18	Low-moderate – High 2.5 – 18.0 50	Moderate – High 40 – 150	Very High 20 - 500
Shape	Elongate	Lobate r belt	Linear belt	Linear belt
Radius/length (km)	100-3000	10-450 0	1-10	1-5
Source area: size	Large Low	Moderate rate	Small	Very small
gradient	Distant	Moderate rate	Moderate	High
distance	Large, mud-rich	Moderate rate	Close	Close
Feeding systems	River delta	Large mixe/narrow river delta ; down-dip c	Narrow shelf	Braid plain
Supply				
Mechanism	Infrequent slumps and slump-initiated low-density turbidity currents. Contour currents	Mainly hig ination low-densit nd high-turbidity c y turbidity its	Collapse of shelfal clastics generating low-efficiency turbidity currents	Frequent mass flow slumps, river generated turbidity currents
Size of flows	Very large	Moderate rate	Small	Small
Channel system	Large, persistent; meandering to straight with well-developed stable levee system	Moderate sple, meandering ht to braided sys lering laterally mi channels with levees	Multiple chutes and poorly developed impersistent channels	Small impersistent chutes
Distal slope/lower fan sediments	Thin, sheet-like flows forming interbedded sands, silts and muds. Coarse intervals forming thin clastic sheets	Mixed-load andy turbidity cu ity current flows formi slumps lobes of interbeddec and muds	Impersistent sand-rich turbidity current flows	Dilute turbidity current flows depositing largely mud and occasional sands
Principal basin plain deposits	Turbidites> hemipelagics	Hemipelagi pelagics turbidites	Hemipelagics	Hemipelagics

MF, mud fan, etc; MR, mud ramp, etc.; MA, mu

Mud-rich depositional systems (<30% sand; Reading and Richards, 1994) are dominated by high-efficiency (after Mutti and Normark, 1987) muddy turbidity currents and flows, in which any available sand will be transported significant distances away from the base of the slope (Fig. 5.14). In comparison, the deposits of mixed mud/sand-rich systems are likely to display characteristics of both high and low efficiency flows, such as deposition of available sand closer to the slope (Emery and Myers, 1996). Both mud-rich and mixed mud/sand-rich systems tend to be characterised by channel-levee complexes (becoming more common in mud-rich systems), chaotic facies indicating mass wasting processes, and smaller depositional or sheet-like basin floor fan lobes (Emery and Myers, 1996; Reading and Richards, 1994). These systems are typical of passive margin basins with wide shelves and high sediment fluxes, and sediment bypassing rather than prograding fan construction (Bouma, 2000b).

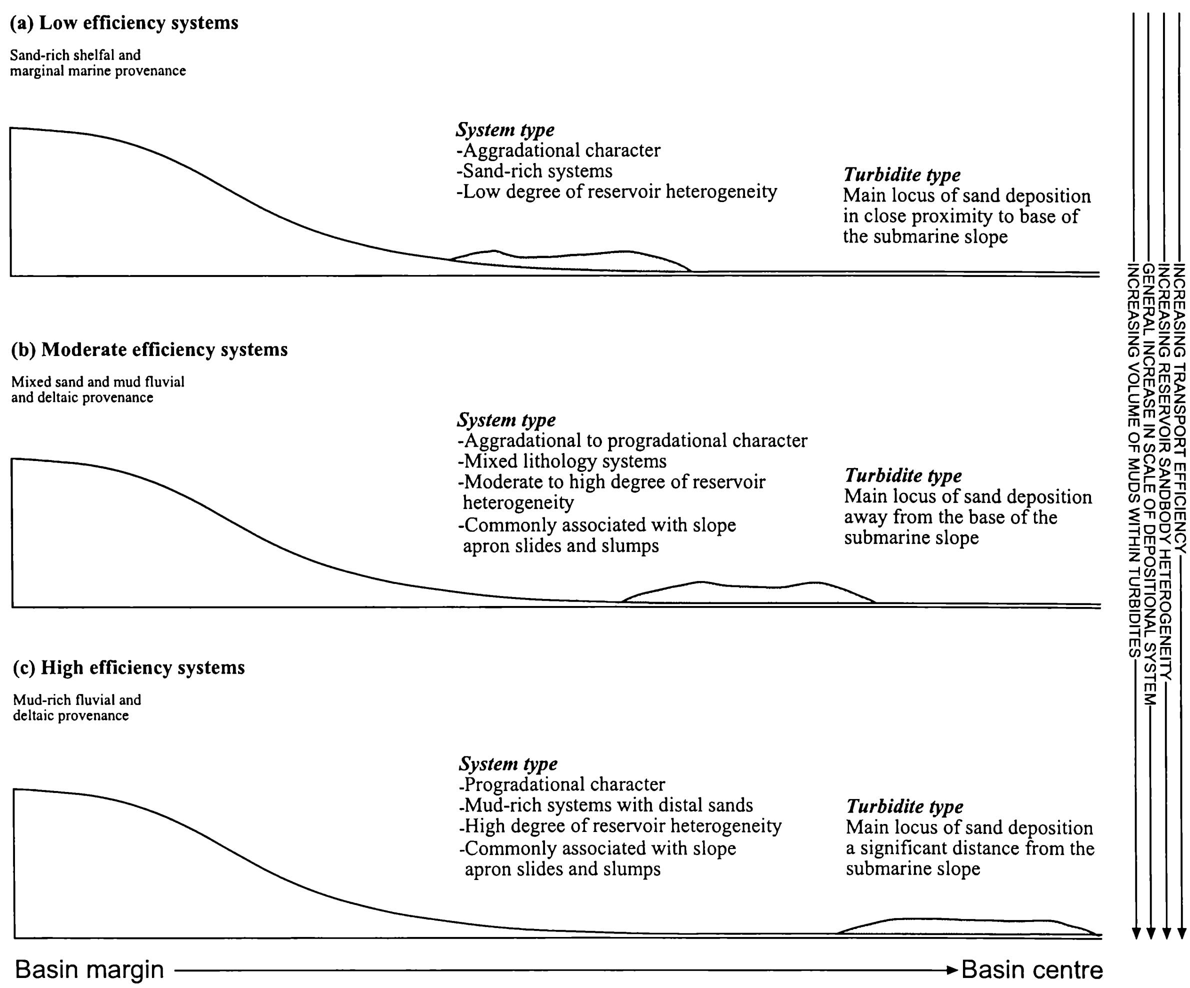


Fig. 5.14: Summary diagram illustrating the concept of transport efficiency, relationships to locus of coarse clastic deposition, and architectural characteristics. From Emery and Myers (1996).

5.6.2 External geometry: the topset-clinoform package

Seismic units that incorporate a foresets component often have a well-developed topset-clinoform package that represents a continuum of shelfal units (topsets) and slope sediments (foresets). The term 'clinoform' is used to describe the more steeply dipping (> 1^{o}) part of the basin margin profile (Emery and Myers, 1996), and a true topset-clinoform package has a distinct offlap break (i.e., break in slope between dipping clinoform reflectors and parallel shelf reflectors; Vail et al., 1991). Bottomsets comprise the lower part of the basin margin profile, and generally contain the deeper water depositional systems. This section gives a brief discussion on the gross external characteristics and geometries of the topset, foreset, and bottomset components of the shelf-slope profile, before examining in more detail the internal reflector characteristics of foreset and bottomset facies.

(i) Topset (shelf) facies

(a) Seismic configuration

Topsets, or shelf facies, are characterised by parallel to divergent, laterally continuous tram-line reflectors that produce a widespread sheet to wedge-shaped distribution on the shelf. Amplitudes of both bounding and internal reflectors vary from high to low, both along a single reflector and within a seismic unit. In some instances, topsets or parts thereof, are characterised by very low to almost reflection free zones. This is attributed to either very thin bed sets (too thin to be resolved by seismic methods) or zones of relative homogeneous lithology. Likewise, termination of a reflection may be due to thinning of the unit, although the unit may still continue below the resolution of the seismic tool (Mitchum et al., 1977b). Topsets are characterised by low gradients (typically less than 0.1^{o}), and on seismic reflection profiles essentially appear flat.

(b) Depositional mechanisms and paleoenvironmental implications

Topsets tend to be characterised by coarser-grained lithologies than foresets, and often contain lenses of shell hash. Wireline facies Sf1 dominate shelfal successions, and if they were to be exposed in outcrop, these topsets would probably resemble the Rangitikei Supergroup shelf equivalents (e.g., Abbott and Carter, 1997; Naish and Kamp, 1997b; McIntyre and Kamp, 1998) or the Matemateaonga Formation discussed in Vonk et al. (2002), which are characterised by a cyclical repetition of shell beds, siltstone, and sandstone lithofacies. In shelf sequences, sea level change is thought to be a controlling factor on the distribution and nature of sediment type (Sangree and Widmier, 1977). Sedimentation on the shelf occurs predominantly during periods of rising and high relative sea level when there is space (accommodation) available for sediment to accumulate. Because seismic reflections tend to parallel stratification surfaces (Mitchum et al., 1977b), the brighter reflectors most likely represent siltstone beds or more condensed units deposited during times of maximum sea level transgression. Topsets often continue laterally into contemporaneous slope facies without disruption, implying high rates of sediment supply, even during periods of relative high sea level.

(ii) Foreset (slope) facies

Progradational clinoform facies

(a) Seismic configuration

Clinoforms within the study area either display classic sigmoid configurations (e.g., seismic units A34-A47; Fig. 5.15a) or more seldomly, sigmoid-oblique configurations (e.g., seismic unit A59, Fig. 5.15b). The slope angles of sigmoidal clinoforms are generally low (a few degrees only) but in places (e.g., in the central part of the P95-seismic grid, and on line AR90-445-103) inclinations are observed to progressively steepen. Where slumping has occurred, progradational slope facies loose their smooth profile, and external bounding reflectors appear more discontinuous and have lower amplitudes. Where clinoforms retain smooth profiles, external bounding reflectors are of moderate to high amplitude and display lateral continuity. This is comparable to equivalent facies described in southern Taranaki Basin by Shell (in Balley, 1987; their sequence 4) in which the actively prograding wedge is characterised by continuous high-amplitude reflections comprising a low-angle downlapping progradational clinoform pattern. Other external geometries include the hummocky relief of seismic unit A54 seen on line P95-158, and an almost linear shape displayed in some thinner units on line AR90-445-103.

Sigmoid clinoforms are characterised by internal reflectors that exhibit concordance and/or convergence with shelfal reflector horizons. Conversely, the less common sigmoid-oblique progradational facies are characterised by reflector termination (toplap) at or near the upper boundary. Both clinoform types display some downlap in the lower part of the slope, and onlap (variously present) in the middle and upper part of the slope (Fig. 5.15a,b). The slope inclination of sigmoid-oblique progradational facies (e.g., seismic unit A59) is steeper than for simple sigmoid facies. This is attributed to a coarser lithology (at Awatea-1 and Te Kumi-1 seismic unit A59 is slightly coarser-grained) and a higher energy environment updip, consistent with a shelf-margin delta fan interpretation.

Progradational facies often form lenses elongated parallel with depositional strike (Sangree and Widmier, 1977), as illustrated in Fig. 5.15a-d. The absence of significant vertical aggradation associated with the clinoforms implies high rates of sedimentation, resulting in rapid progradation (e.g., Adams et al., 1998).

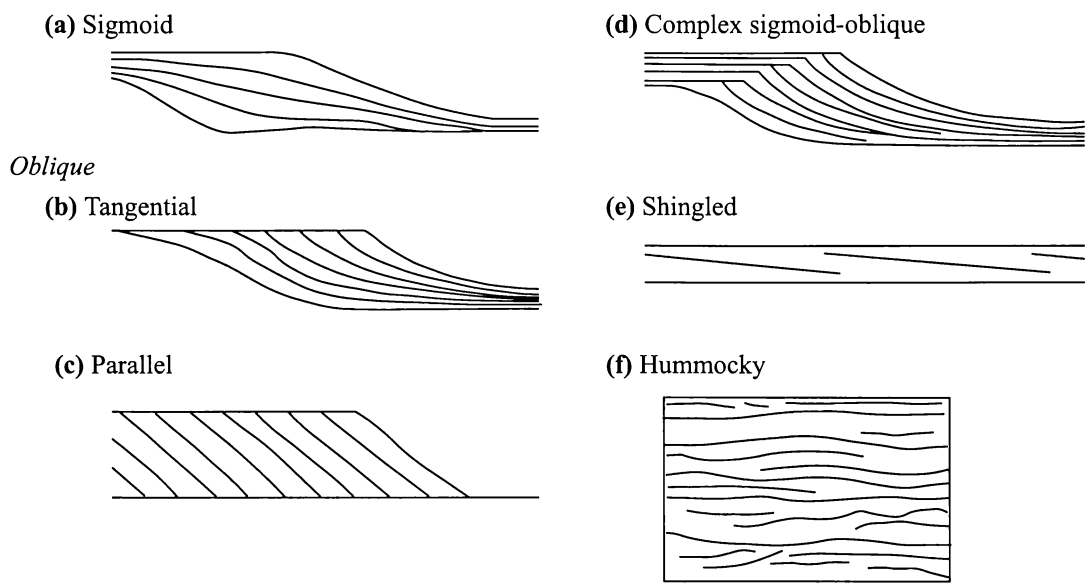


Fig. 5.15: Typical clinoform reflector configurations. After Mitchum et al. (1977b).

While sigmoid and sigmoid-oblique are the dominant clinoform patterns observed within the Giant Foresets Formation, a third type of progradational clinoform is also recognised. This clinoform pattern, referred to as 'shingled', is characterised by offlapping reflector configurations (Fig. 5.15e). It is observed higher in the succession, within topset (shelf) facies, and characterises several seismic units, including A67 and A70. Although it is essentially a topsets feature, it is included here because of the progradational nature indicated. These progradational-shingled configurations occur when the shoreline does not drop near the shelf break during regression, or as a result of an abundant sediment supply during high relative sea level conditions.

(b) Depositional mechanisms and paleoenvironmental implications

Progradational clinoforms form in response to basinward migration of the shelf margin, reflecting changes in relative sea level and/or depositional flux (section 5.8). Progradational facies are generally dominated by fine-grained lithologies and low energy (turbiditic or suspension sedimentation) deposition (Sangree and Widmier, 1977; Jervey, 1988), with slope angle and curvature strongly influenced by sediment 'caliber' (Emery and Myers, 1996). While steeper slopes in clastic systems tend to be composed of coarser lithologies, or are zones of sediment bypass (Emery and Myers, 1996; Pirmez et al., 1998), Adams et al. (1998) suggest that submarine slope curvature from passive continental margins, in areas of high sediment supply and active progradation, is controlled by the exponential decay of transport capacity with increasing distance from the shelf break. Further to this, Adams and Schlager (2000) recognised three slope types that could be represented by simple equations. A gaussian distribution could describe fifty percent of slope profiles studied, including sigmoidal morphologies, with mud-dominated slopes having lower slope angles than sand-dominated slopes.

Comparisons with prograding clinoforms of the Giant Foresets Formation reveal a dominance of Gaussian slope profiles associated with the foreset front. While these profiles are associated with lower sedimentation rates (compared to linear or exponential curves), Adams and Schlager (2000) indicate that modifying effects, such as a fluctuating base level, sediment creep, slumping, turbidity currents and debris flows, as well as the influence of oceanic currents, contribute to the predominance of Gaussian curves observed, and maintain the inclination of the depositional slope in the range of 1°-5°. Modifying effects are strongly emphasised by the gradual increase in inclination (observed in both the Western Stable Platform and Northern Graben) as clinoforms prograde into deeper water, resulting in oversteepening and slope failure.

Degradational foresets

(a) Seismic configuration

Degradational foresets (seismic units A71 to A73) incise deeply into, and abruptly terminate, underlying seismic units. These units (particularly A72 and A73) are extremely thick compared to progradational foresets, and at Ariki-1 and Tangaroa-1 comprise a significant proportion of the Nukumaruan and younger successions (see Appendix 6).

Bounding surfaces of each unit vary from moderate amplitude, relatively continuous, to low amplitude and discontinuous. Internally, these units are characterised by numerous multiples, and highly erratic, disrupted and discordant or chaotic reflector configurations basinward of the preceding shelf break, becoming more continuous and parallel up through the unit. Landward, reflectors are more continuous, and although the units thin rapidly landward of the shelf break, are laterally extensive. The upper boundary of seismic units A72 and A73 are ravinement surfaces, characterised by incision of small to large scale channels. On line P95-158 (Enclosure 3) the large channel incision into seismic unit A73 to the west of Tangaroa-1 is overlain by hemipelagic drape, indicating that channel incision in this area has ceased. This in turn suggests that sediment is presently bypassing this area.

(b) Depositional mechanisms and paleoenvironmental implications

While a smooth upper to mid slope profile reflects progradational outbuilding, the highly dissected morphology of the degradational foresets is the physical manifestation of active degradation due to oversteepening of the shelf margin. Beggs (1990) states that beyond a critical angle, empirically determined to be about 3° for the Giant Foresets Formation, slope instability leads to mass failure and secondary down-slope movement of sediment. This results in formation of the degradational facies on the slope and base of slope. Lithology may also play an important role in determining the resultant morphology of shelf-slope facies. Coarser lithologies produce steeper slopes (Emery and Myers, 1996); wireline and well cuttings demonstrate that sediments are somewhat coarser and shellier through degradational facies than progradational facies. This in part reflects higher energy shelfal conditions (water depths shallowed considerably through the Nukumaruan), but may also reflect an additional or stronger influence from a secondary sediment source.

As a smooth upper to mid slope profile reflects progradational development, slope morphology can be used to infer the age of most recent progradation. This study concurs with Beggs (1990), placing the currently prograding sector to the north and west of the P95-seismic grid. Bathymetric charts (e.g., CANZ, 1997) show that south of Tane-1 to about Te Kumi-1, the upper slope bathymetric contours are highly serrate, whereas north of Te Kumi-1, the contours are much smoother.

(iii) Basin floor facies

(a) Seismic configuration

The lower part of the basin margin profile, where clinoforms pass laterally into flatter lying reflections, are termed bottomsets (Vail et al., 1991). In most cases within the study area, bounding reflectors of bottomset facies are concordant with clinoform reflectors. However, in some cases (e.g., westernmost end of line P95-158) the bottomset reflectors are not in depositional continuity with the clinoforms, and instead form separate packages onlapping the clinoform front. These latter bottomsets often display a different internal geometry and reflector configuration than the former, and are further discussed in section 5.6.3(iii).

External geometries and bounding reflector characteristics of bottomset facies vary with internal reflector characteristics and relationship to the slope portion of the overall profile. Where bounding reflectors are concordant with clinoform reflectors, bottomsets tend to display gently mounded to gently basinward-dipping external forms. In cases where bottomsets are disconnected from foresets (e.g., seismic unit A8), external geometries are more wedge-shaped in cross-sectional view, and internal reflector patterns tend to be more chaotic and discontinuous.

Older seismic units (seismic units A1 – A19) are only represented by bottomsets (presumably the slope and shelfal components are much further to the south and/or east). These units are dominated by tabular geometries, becoming broadly mounded when associated with the Mangaa Formation. Internal reflector configurations are dominated by parallel to sub-parallel, relatively continuous and moderate to high amplitude reflector horizons, although some units degrade to chaotic and hummocky, discontinuous and low amplitude reflectors (e.g., seismic unit A19 to the west of Tangaroa-1).

(b) Depositional mechanisms and paleoenvironmental implications

Bottomset deposits represent deepwater depositional systems, and thus form as a result of a number of depositional mechanisms. These include high-density mass flows (debris flows, high-density turbidity currents) and pelagic/hemipelagic suspension sedimentation. The specific mechanisms and types of depositional environments encountered in the study area are discussed further in section 5.6.3.

5.6.3 Internal reflector configurations

(i) Basin floor fans

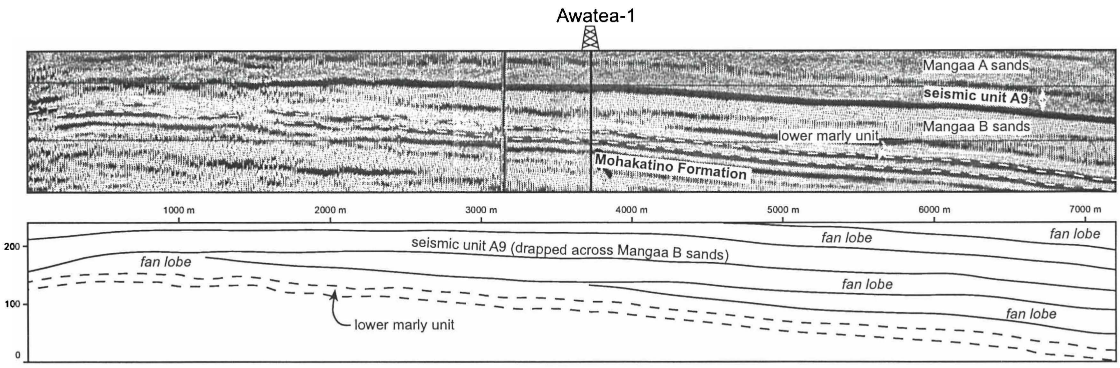
(a) Seismic configuration

Basin floor fans are variously termed submarine fans (modern analogues), lowstand fans (sequence stratigraphic models, e.g., Vail, 1987), turbidite systems (Mutti and Normark, 1987), suprafan lobes (Shanmugam and Moiola, 1991, later abandoned by Shanmugam, 2000), depositional lobes (Reading and Richards, 1994), and lower fans (Bouma, 2000b). They are identified seismically by their morphology and by position in the overall paleogeographic setting (Mitchum, 1985). Several distinct types of basin floor fan architectures have been recognised in this study (Table 5.3).

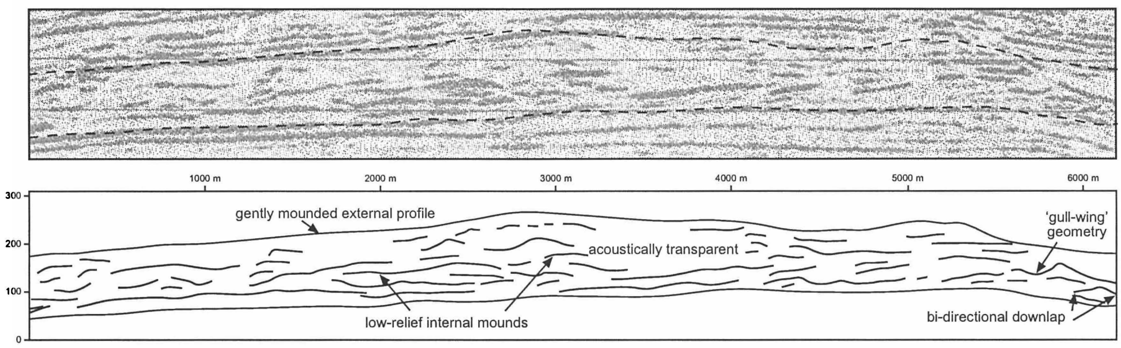
Table 5.3: Characteristics of three types of basin floor fans.

	External geometry and bounding reflectors	Internal reflector configuration	Aerial extent	Example and representative seismic line
Type 1	Gently mounded, sheet-like to wedge shaped geometry Continuous, bright, high amplitude reflectors	Parallel to subparallel, moderate to high amplitude and continuous, becoming less so towards the outer fringe	Moderate-large, several 10's of km's Fault controlled, several lobes identified	Mangaa Formation (P95-138)
Type 2	(a) Gently mounded High to moderate amplitude reflectors	(a) Low-relief mounds, and/or gull-wings, bi-directional downlap, often discontinuous, variable amplitude, occasionally acoustically transparent	Restricted, several km's Elongate lobes	Seismic unit A41, A54 (P95-158)
	(b) Planar geometry	(b) Parallel and continuous, moderate amplitude, or reflection free	Grades laterally into pelagic/hemipelagic basinal facies. Extent uncertain	Seismic units A34, A39,
Type 3	Distinctly mounded geometry Bold, moderate to high amplitude reflectors	Numerous parallel-subparallel, moderate amplitude and continuous reflectors downlapping lower boundary	Large, several 10's of km's Roughly circular to elongate lobes	Seismic unit B6 (NM 17)

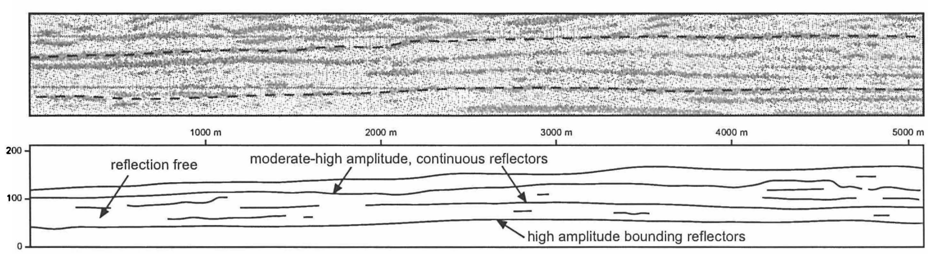
Because it is difficult to delineate mud-rich fans from sand-rich fans solely on seismic criteria, the three types of basin floor fan are also distinguished based on wireline characteristics and interpreted lithologies. Type-1 fans (of which the Mangaa Formation appears to be the only example in the study area) display blocky SP and resistivity motifs (Bff1 wireline facies), and are dominated by massive sandstone beds separated by thinner mudstone beds (e.g., Mangaa-1, Appendix 2, 6). This heterolithic character is reflected in the nature of internal reflectors (Fig. 5.16a). Lithologies undoubtedly become siltier and more homogeneous towards the fringes of depositional lobes, reflected by a lateral change in reflector characteristics. Type-2a fans are characterised by finer-grained fining- and coarsening-upward (Sf2) wireline facies, with little evidence of coarse sandy lithologies, even at the apex of the external mounded forms (refer to Tangaroa-1 well site, Enclosure 3). Type 2b fans display subdued responses, indicative of muddy turbidite and hemipelagic accumulation.



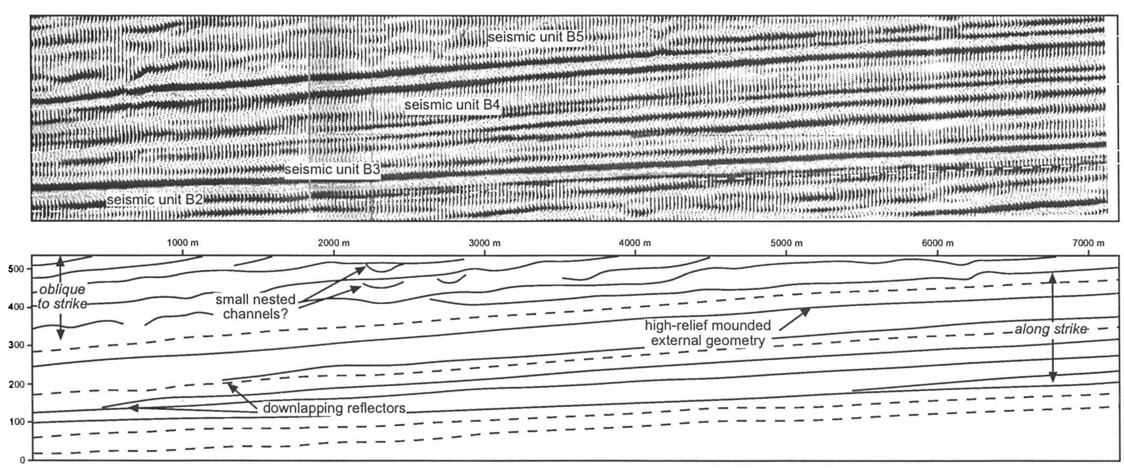
(a) Type 1 (Mangaa Formation, seismic reflection profile P95-115).



(b) Type 2a (seismic unit A41, seismic reflection profile P95-158).



(c) Type 2b (seismic unit A34, seismic reflection profile P95-158). NB: aerial extent much greater than illustrated.



(d) Type 3, (seismic units B4-B6, western end of seismic reflection profile NM16). See also Fig. 5.10.

Fig. 5.16: Three main depositional styles and seismic expression of basin floor fans within northern Taranaki Basin. Note vertical scale for all figures is in msecs (TWT), and are scaled at 1.5x original.

Type-3 fans are the largest of all fan types, and are often characterised by much muddier lithologies than overlying topset beds (e.g., Arawa-1, Appendix 2, 6), although GR logs for both Arawa-1 and Taimana-1 display abundant small-scale fining- and coarsening-upward log motifs (Sf2).

Type-2a fans display the most complex internal geometries and seismic reflector configurations (Fig. 5.16b). External mounded forms are characterised by numerous smaller, high-amplitude overlapping mounds, which are further characterised by parallel reflector sets (e.g., seismic unit A54), interspersed with zones that are almost acoustically transparent. The transparent nature of some fans is often indicative of basin floor sands, but can also indicate relative homogeneity, and thus probably reflects the muddier nature of the Giant Forests Formation. Each external mounded reflector defines internal lobes and depositional axes within the fan. Internal constructional mounds commonly fill depositional lows between older lobes. They are delineated by both their convex-upward to hummocky form or gull-wing geometries, and by bi-directional downlap. Internal mounds occur in the order of 10-30 msecs in relief, and a few hundred metres in length. Downdip, mounded packages grade into parallel-subparallel basin floor facies, and occasionally into more chaotic facies (generally where an obvious change in gradient occurs). Updip, reflectors become more concordant with slope complexes, or onlap younger reflectors along the edge of the fan complex. Type 2b (Fig. 5.16c) fans are typical of muddy depositional environments, characterised by continuous, high-amplitude parallel reflector sets (e.g., A34) or reflection-free zones (e.g., A47).

Not all seismic units have an associated fan component, and with few wells it is difficult to determine the lateral facies changes from slope to basin floor. The lack of large fan mounds typical of basin floor models (e.g., Posamentier et al., 1991) suggests that the basin floor was fed via a number of sources (line source), creating an elongate prism of slope sediment (slope apron; Galloway, 1998), and Type-2 depositional lobes. Conversely, the external geometry and internal reflector configurations of both Type-1 and Type 3 (Fig. 5.16d) fans are suggestive of a point-source feeder system.

(b) Depositional mechanisms and paleoenvironmental implications

Basin-floor fans are most often indicative of sedimentation during the early phase of relative sea level fall (Van Wagoner et al., 1988). During this phase, channel incision related to base level lowering results in sediment by-pass of the shelf system (Mitchum, 1985), and the focus of deposition is shifted from the nearshore regions to the basin. Although fan sedimentation can occur at any stage of a sea level cycle (Stelting et al., 2000), large influxes of clastic material are more likely to reach the outer shelf as the shoreline regresses basinward towards the shelf break (Sangree and Widmier, 1977; Posamentier et al., 1991). Depositional processes contributing to fan formation are thought to be controlled by turbidity currents, although there is some debate as to whether these are in fact dominant, especially in sand-rich fans (e.g., Shanmugam et al., 1995; Shanmugam, 2000; Stow and Mayall, 2000).

While fans can have any net-to-gross sand percentage (Bouma, 2000a), sand-dominated Type-1 fans conform more readily to models of basin floor fan development cited in the literature (e.g., Mitchum, 1985; Posamentier et al., 1991). The mounded-relief and sheet-like fan lobes of the Mangaa Formation (Fig. 5.16a) appear to be detached from the concomitant slope, display good lateral communication (at least between Mangaa-1 and Awatea-1), and are reasonably extensive. Deepwater fan models position fan development at or near the mouths of submarine canyons (Shanmugam and Moiola, 1991), although no channel or canyon has yet been tied to the Mangaa fan system. However, Mitchum (1985) states that some channels may be poorly developed or far removed from the fan.

The change from Type-1 (Mangaa Formation sand-dominated depositional fan lobes) to Type-2a and b (mixed mud/sand- and mud-dominated) fans in the northern part of the study area suggests a change in source area and/or supply system. Sand-rich point-source fans are inferred to have been sourced from a nearby hinterland (Emery and Myers, 1996), which suggests that, rather than being transported long distances via some as yet identified channel or canyon system as suggested above, the Mangaa Formation was in fact sourced from a closer hinterland (i.e., the proto-North Island; recall that Forder and Sissons, 1992 suggest that Mangaa Formation sediments are co-dominated by a Murihiku Supergroup provenance). This source area may have been abruptly shut off due to continuing uplift across the Turi Fault Zone at about the same time that sediment derived from the emerging Southern Alps moved across the widening shelf into the northern part of the study area. The change from a low efficiency, sand-rich system, to a high efficiency mud- mixed mud/sand-rich system was translated into the range of submarine geomorphologies and stratal elements now observed on seismic reflection profiles.

(ii) Channel-levee complexes

(a) Seismic configuration

Channel-levee complexes occur upslope of basin floor fans, and are variously described in the literature as slope fans (e.g., Van Wagoner et al., 1988; Posamentier et al., 1991) and upper fans (e.g., Mitchum, 1985). Their accumulation may be coeval with the basin floor fan, or in place of one another (Van Wagoner et al., 1988; Emery and Myers, 1996). In seismic sequence terminology, they comprise a portion of the lowstand systems tract characterised by turbidite and debris flow deposition on the middle or base of slope (Van Wagoner et al., 1988).

Channel-levee complexes are recognised on seismic reflection profiles by moderate to low amplitude mounded or gull-wing (concave up-ward) reflector configurations, bi-directional downlap and bow-tie diffractions at channel edges (Mitchum, 1985). Within the progradational foreset facies of the Giant Foresets Formation, few classic gull-wing configurations are observed – where they are obvious, they tend to be poorly developed with diffuse reflectors downlapping away from the channel axis. Most channel-levee systems are dominated by small-scale, irregular hummocky to mounded profiles, characterised by moderate amplitude reflectors that become more diffuse away from channel margins (Fig. 5.17). Channel-levee complexes occur as single channel and levee couplets, or more seldomly, as smaller-scale nested channels and vertically stacked levees. Overbank deposits, the lateral equivalent of channel-levees, are delineated by their moderate continuity and generally higher amplitudes. Channel fill is characterised by acoustic transparency, and is probably mud-dominated.

The relationship of seismic unit A54 to seismic unit A55 (P95-158) is diagnostic of large channel-levee margins, including hummocky geomorphology, roll-over of internal reflectors towards the deeply incised channel, and divergence away from the channel margins. Channel margins of this scale are not observed elsewhere in the study region.

For the most part, channel-levee complexes within particular seismic units are coeval with the basin floor fan, but occur higher up the slope; these complexes are typically overlain by thinner intervals characterised by moderate to high, relatively continuous, thin reflectors that downlap the complexes and either onlap the upper seismic unit boundary, or roll-over into concordant topset reflectors. The sediments comprising these intervals are interpreted to represent turbidite strata deposited on the slope environment during periods of rising and high relative sea level (lowstand wedge of Van Wagoner et al., 1988), as energy of the depositional system waned.

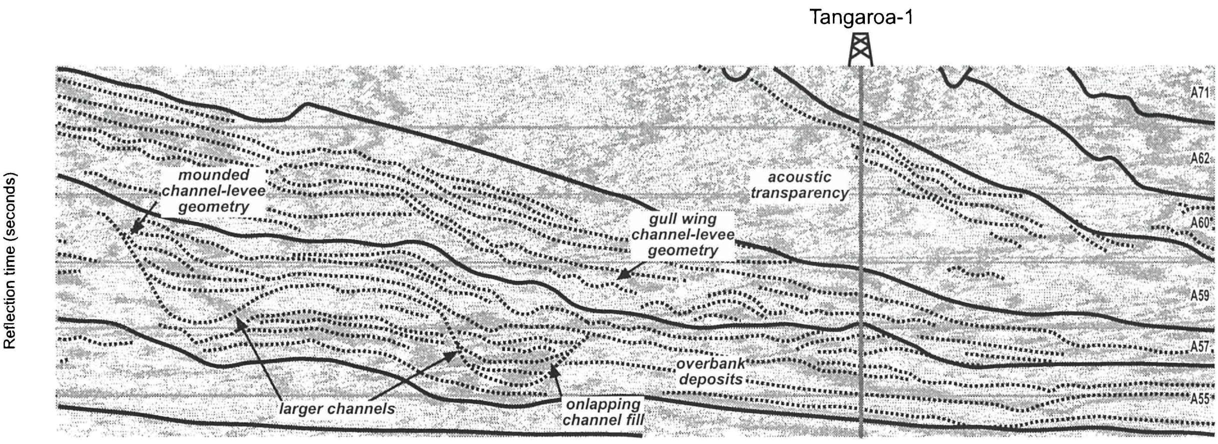


Fig. 5.17: Channel-levee complexes, seismic reflection profile P95-103. Channel-levee complexes are the dominant slope architectural elements, and are often found at the base of the slope in conjunction with basin floor fan facies. Note onlapping channel fill, mounded or gull wing levee geometry, and parallel-subparallel overbank deposits associated with larger channels. Acoustic transparency is a common feature of channel-levee systems. Vertical scale 1.5x original.

Occasionally, shelf margin delta fans, part of the lowstand prograding wedge, occur (e.g., seismic unit A59). The complex-sigmoid oblique clinoforms that characterise this unit are typical of deposition in a shelf margin deltaic environment (Sangree and Widmier, 1977). Toplap truncation indicates that relative sea level dropped low enough that strata were not able to extend updip (Mitchum et al., 1977b).

(b) Depositional mechanisms and paleoenvironmental implications

Channel-levee complexes, associated channel-fill and overbank deposits, and shelf-edge delta fans are considered to be components of the lowstand prograding wedge, deposited during static and slowly rising relative sea level conditions when sediment influx is still relatively voluminous (Sangree and Widmier, 1977; Posamentier et al., 1991). Channels form the conduits for terrigenous sediment transport from shelf to basin floor, developing where transporting flows enter the base of the slope as the result of an efficient bypassing system (Bouma, 2000b). Levees and overbank deposits are built up when turbidity currents and other mass flow processes overtop or breach existing levees and deposit sediment beyond the channel (Kirschner and Bouma, 2000). Consequently, channels tend to be sand prone (and are often targets in the oil exploration industry), while channel-levee and overbank deposits are mostly silty to muddy.

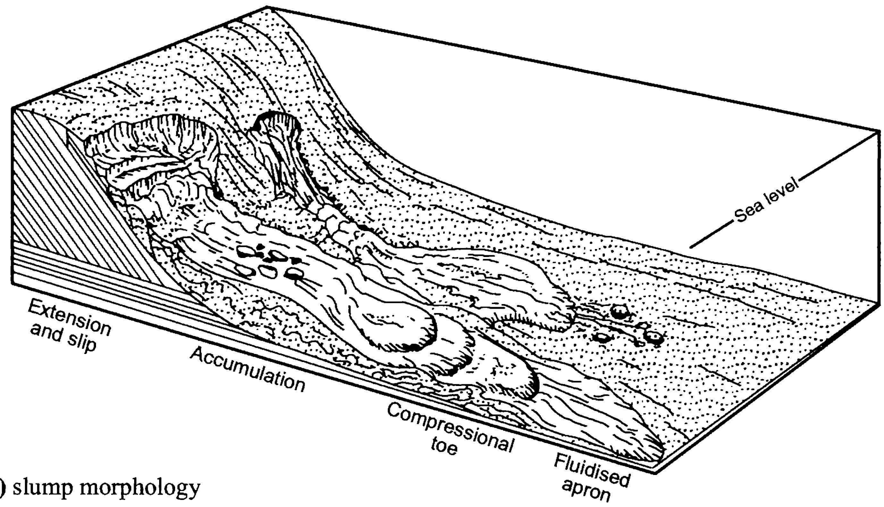
In sandy systems, channel-levee reflectors often change from moderate to high-amplitude reflectors at the base of the levee, to low amplitude near the top (e.g., Weimer, 1990). This is interpreted to reflect a decrease in grain size and bed thickness of the levee system. This differs from channel-levee systems in the Giant Foresets Formation, predominantly due to the fact that channel-levees are more often than not only represented by a single reflector. Stacked complexes (e.g., seismic units B13-B16, Wainui-1, Appendix 6) do display an upward decrease in grain size (wireline log motifs), but overall, reflect a dominance of fine-grained lithologies. The lithological homogeneity of sediments provides an explanation for the very low amplitudes observed on seismic reflection profiles, as levee and channel elements may not be imaged seismically when their lithologic character is similar to the background hemipelagic sediments (Emery and Myers, 1996).

Overbank deposits, being more often than not represented by subparallel-parallel, moderate amplitude and relatively continuous reflectors suggest deposition via turbidity currents derived from a channel. Occasionally, overbank deposits are more disrupted and discontinuous in nature, which indicates derivation from small-scale slides or debris flows. Seismic resolution is not great enough to reveal the scale of detail observed in outcrop studies of channel-levee systems (e.g., Coleman et al., 2000b; Cronin et al., 2000); direct observation of the spatial and temporal relationship between successive channel-fills and channel margins suggests that levees are temporally unrelated to channel-fills, whereas overbank deposits can be correlated with post-levee formation channel-fill (Cronin et al., 2000).

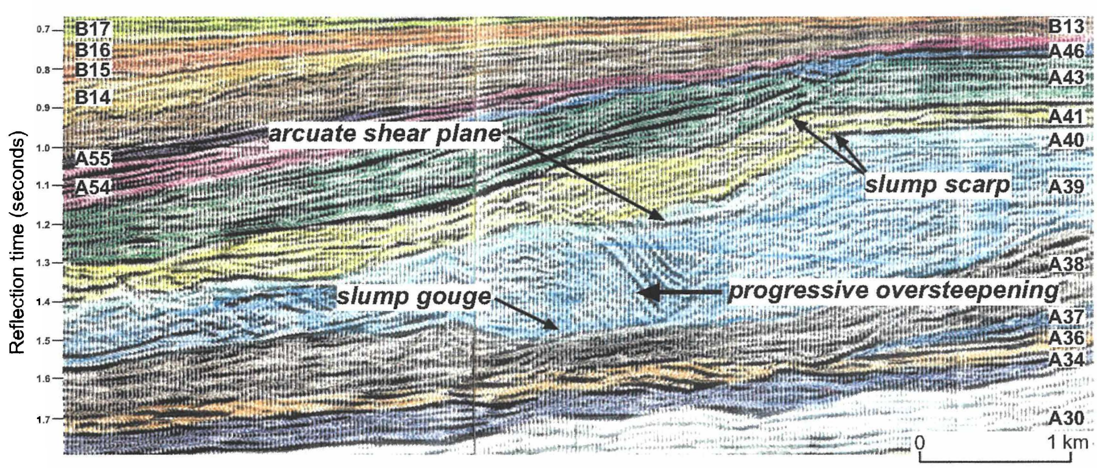
(iii) Mass wasting

(a) Seismic configuration

Mass-wasting processes such as slumping, sliding and debris flow emplacement occur on a number of different scales within the study area. The largest and most obvious is the slumping associated with seismic units A72 and A73 (degradational facies). Seismic reflectors within these units are variable, displaying a range of configurations from reflection free or low amplitude discontinuous reflectors (particularly the lower part of seismic unit A72), to moderate to high amplitude relatively continuous, parallel to subparallel reflectors. In some places discrete and coherent blocks that still retain their original tramline reflector configurations are evident. Internal onlap configurations, while rare, indicate back-building (up towards the shelf) once slumping had ceased. Prominent scarps and basal gouges associated with unit A72 (and also seismic unit A60 – see line P95-158, Enclosure 3) attest to the erosive nature of these units. Bounding reflector characteristics are variable, ranging from low to high amplitude, and discontinuous to relatively continuous. While it is not possible to interpret the types of internal slump-features that might be present in these (and other) units, they are probably somewhat analogous to those displayed in the slumped Tawariki Formation interval that crops-out on the northern Taranaki coastline (including isoclinal folds and rafted blocks of finer or coarser lithologies; see e.g., King et al., 1993; Wilson, 1994; Hansen, 1996).



(a) slump morphology



(b) Part of the Manganpanian-Early Nukumaruan slumped interval.

Fig. 5.18: Seismic expression of slumped units. (a) Morphology and depositional architecture of a large slump (from Galloway, 1998, after Galloway and Hobday, 1996); (b) Mangapanian - Early Nukumaruan slumped interval, seismic reflection profile AR90-445-103. Note the sharp slump scarps, curved basal slump gouges, and internally-chaotic seismic character. Mounded external profiles associated with a compressional toe are observed associated with seismic units A38 and A41 (see Fig. 5.10).

A broad stratigraphic interval of smaller-scale slumped units occurs within the foresetted part of the Giant Foresets Formation. This band includes seismic units A36 to A40, and parts of A43 and A41, and is typified by variable internal characteristics, including acoustically transparent, chaotic and discontinuous to subparallel and continuous, moderate amplitude reflector configurations in the mid to upper slope regions, becoming more continuous and concordant in the lower slope. Bounding reflectors are often low amplitude and hummocky, and the lower boundary may display arcuate shear planes, indicative of rotational slumping (Fig. 5.18). These units may display external mounded forms rather than sigmoidal forms (e.g., seismic unit A40, line P95-158) and are often quite extensive. In places, the upper part of slumped units are characterised by extensive stacked channel-levee systems in the mid to upper slope region (e.g., A40). In the northern part of the study area, another stratigraphic slumped interval, with similar internal and external characteristics, has been identified. This younger interval (Fig. 5.19), encompassing seismic units A48 to A51, has a more restricted distribution than the previous slumped interval. The distribution pattern suggests that these units occurred as a result of movement on nearby Kahawai Fault. Small-scale curvilinear extensional faults are also associated with local slumping (e.g., Mangaa-1, Appendix 6), and slumping may be observed at the edges of some channel margins.

A distinctly different seismic configuration is observed in the lower slope/toe of slope portion of seismic unit A55, clearly evident on both P95-158, and P95-103 (Enclosure 3). The basal part of this seismic unit displays the morphology of a relatively large single event slump, with an acoustically transparent internal character with few or no internal reflectors, a mounded external form, and a blunt compressional toe (e.g., Fig. 5.18a). The uneven topography initially created by this slump was later in-filled by a channel-levee complex. On isopach maps (Fig. 5.12d), this slump is defined by a tongue of thicker deposition to the west of the main depositional loci of seismic unit A55.

High-density debris flows are distinguished from slumps by several criteria. While characterised by similar internal reflector configurations, they are only recognised basinward of the toe of the slope and are often not concordant with slope clinoforms, are aerially restricted, and display wedge-shaped geometries in cross-section. Within the study area, debris flows are predominantly recognised in the deeper parts of the basin (west and north of Ariki-1 and Wainui-1, also observed on the western end of P95-158, Enclosure 3) often some distance from the slope. Eschard (2001) has demonstrated that clastic material can be transported hundreds of kilometres into the pelagic domain.

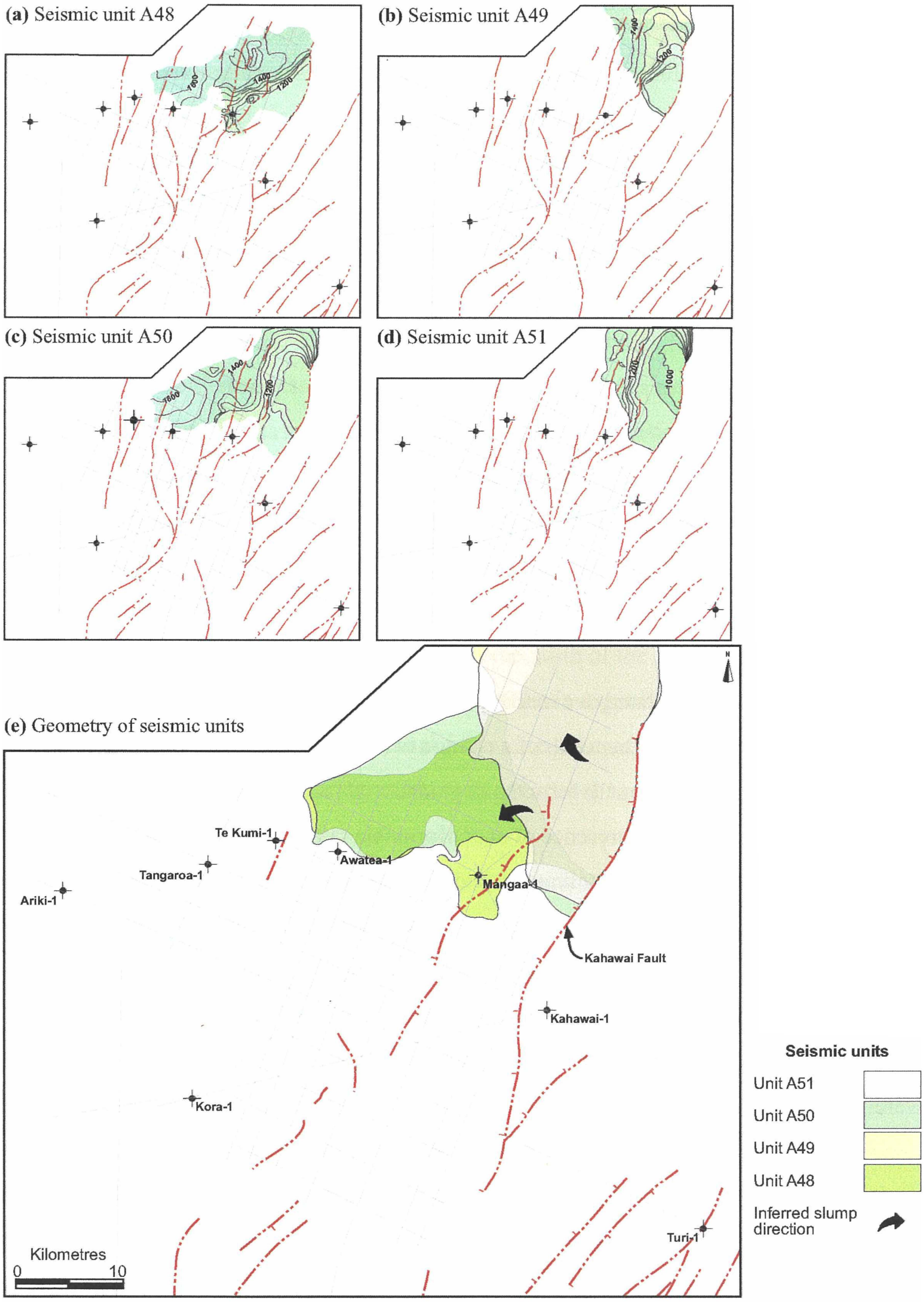


Fig. 5.19: Seismic units A48 - A51 (a-d), in the northeastern quadrant of the study area are characterised by highly disrupted, discordant and chaotic reflectors. While these may be the distal toes of progradational wedges or submarine fans, it is more likely from their seismic character, position and overall geometry (e) that they are slumped units, possibly related to movement on Kahawai Fault to the east. Depth to structure is in TWT (msecs).

(b) Depositional mechanisms and paleoenvironmental implications

Slumps, debris flows and other catastrophic failure events are initiated by a complex interaction of several variables, including active tectonics and faulting, water level changes, slope morphology, and sediment accumulation (Coleman and Prior, 1988; Pratson, 2001). The greatest numbers of failures occur on the slope because this is where inclinations tend to be greatest, and sedimentation rates are often very high (Pratson, 2001). Subaqueous mass wasting can constitute some of the largest slope movements on earth, but information about the mechanisms of movement is poor because they must usually be inferred after the event (Coleman and Prior, 1988). However, various studies (e.g., Elverhøi et al., 1997) have shown that mass wasting processes are often triggered by instability of the shelf-slope margin during relative sea level fall. As sea level is lowering, sediment may be rapidly transported basinward. This causes sediment overloading, which in turn increases pore-fluid pressures, leading to a reduction in sediment strength (Coleman and Prior, 1988). This may eventually lead to oversteepening of the shelf margin, and result in catastrophic failure, and the transport of sediment deep into the ocean basin.

Slope instability, possibly related to a marked decrease in water depths during the Nukumaruan (Chapter 6) and also possibly the result of a change in sediment character, led to repeated mass failure and slumping of sediment to form the degradational facies of seismic units A72 and A73 (and possibly also A60?). The presence of unconformable rafts or glide blocks within the lower part of seismic unit A72, and continuous undeformed reflectors in the upper part of each slumped unit is suggestive of a translational-slide component (Emery and Myers, 1996).

(iv) Submarine channels

(a) Seismic configuration

Several channel systems have been mapped in the Taranaki Basin, such as the top Miocene channel network of Thrasher and Cahill (1990), as well as being imaged in 3-dimensional seismic reflection profiles of the Moki Sandstone in southern Taranaki Basin (Bussell, 1994). Within the study region, aside from the channels associated with channel-levee systems, several larger-scale channel networks have been identified and mapped at a number of stratigraphic intervals within the Giant Foresets succession (Fig. 5.20).

Channels are delineated by truncation of underlying strata, and by distinctive channel-fill configurations (refer to Appendix 5b for configurations). Several conspicuous channel systems, varying in width and depth, are identified on seismic line P95-158 (Enclosure 3). Most obvious is a large channel observed in the central part of this line. This U-shaped channel is up to 300 msecs in depth, incising deeply into the underlying seismic units, and up to 7 km in width. Channel-fill displays a complex seismic configuration, with mounded onlap fill in the lower part (suggestive of slumping of the channel walls, or inter-channel fan-type deposition; Mitchum et al., 1977b; Brown and Fisher, 1980), and divergent or migrating fill in the upper part (caused by differential compaction of underlying channel-fill or by deposition from turbidity currents or mass flows).

More commonly, however, channels are of a much smaller scale (hundreds of metres wide and less than 100 msecs in depth), such as observed on Enclosure 3 (P95-158) between P95-155 and P95-161 line intersects. Channels of similar dimensions occur at various stratigraphic intervals, though most occur in Late Nukumaruan-aged topsets strata. These younger shelf channel systems can be mapped across the central part of the study area (Fig. 5.20), but are less frequent to the north of Mangaa-1, and south of line P95-198. Stagpoole (1997) also observed that channel deposits were uncommon in the region to the north of the study area. The meandering nature of these younger channels is an indication of the low gradient associated with the shelf (Pickering et al., 1995). Associated channel fill is often characterised by very low amplitude reflections, with onlap fill configurations indicating sea level rise (Brown and Fisher, 1980).

The youngest channels in the study area occur in association with seismic unit A73. The channel incised into this unit west of Tangaroa-1 (see Enclosure 3) displays similar characteristics to older infilled channels, with a broadly U-shaped profile, ancillary channels immediately to the west, and relatively steep sides. This channel is overlain by pelagic drape, indicating that the channel is no longer a conduit for sediment transport.

Often, reactivation of older channel systems is evident on seismic reflection profiles, with subsequent channels preferentially incising into previously deposited channel fill (e.g., between lines P95-131 and P95-001 on line P95-158, Enclosure 3). In many cases, two or more stacked channels are associated with small-scale faults, which suggests that activation of these faults may have created a zone of weakness along which subsequent channels incised. Syn-sedimentary faulting can in fact strongly control channel courses (Pickering et al., 1995). In other instances, channels display a stacked-offset nature, similar to that seen within the Mount Messenger Formation cropping out along the Taranaki coastline, which implies that the channel system in place at this time migrated backwards and forwards much as a terrestrial meandering or braided stream network might do.

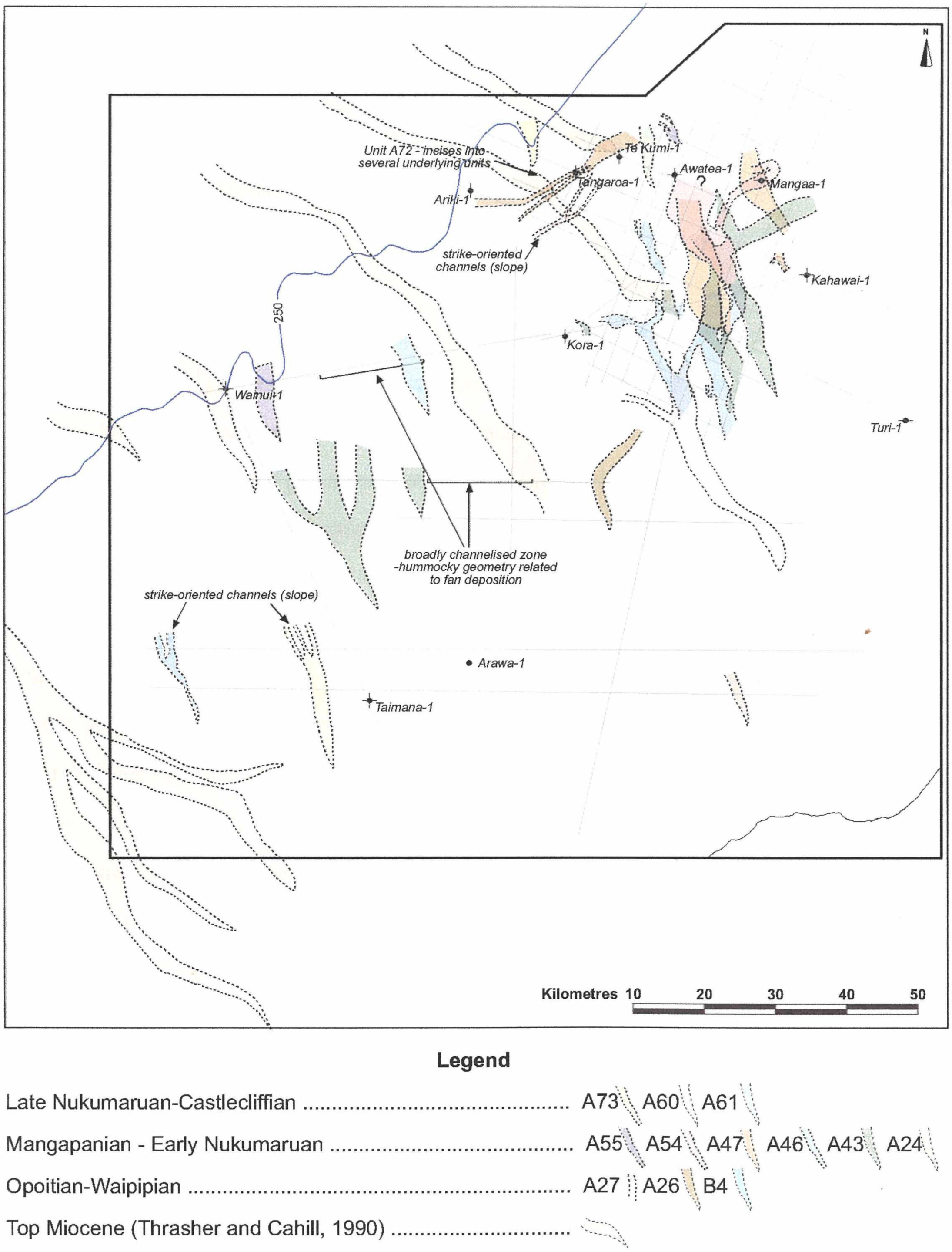


Fig. 5.20: Major Plio-Pleistocene channel systems, northern Taranaki Basin. Top Miocene channels are after Thrasher and Cahill (1990). Larger channel systems often trend northwest-southeast, and form broadly channelised zones. Most of the large channel systems are associated with latest Opoitian to Early Nukumaruan strata, while smaller systems are associated with Late Nukumaruan and Castlecliffian strata. The colours of channels represent the colour of the associated seismic unit into which the channels are incised.

Apart from the large channel evidenced on P95-158, younger channels in general are shallower and narrower than channels lower in the succession. No obvious channels occur within the slumped interval discussed in the previous section (seismic units A36 to A40). This is typical of mud-rich systems, where distribution patterns do not appear to be strongly controlled by deeply incised canyons or channels (Stelting et al., 2000).

(b) Depositional mechanisms and paleoenvironmental implications

Channels form during lowering and low sea level, with headward erosion into the existing marine topography aided by submarine failure of previously deposited unstable lowstand deposits. Many variables, such as hydrodynamic conditions, shelf width and gradient, rates of eustasy, tectonics, and migration of sediment sources act to create a diversity of submarine canyons and channel types (Farre et al., 1983; May et al., 1983; Lindseth and Beraldo, 1985). As channels develop, they become more sinuous, particularly where gradients are lower (Pickering et al., 1995).

Mud-rich and mud/sand-rich depositional systems are dominated by well-developed (though low-amplitude) channel-levee systems, which form the main conduits through which sediment is transported from the upper shelf/upper slope to the basin floor (Emery and Myers, 1996). In mud-dominated systems, large channels with a well-developed levee system (e.g., like that observed on line P95-158) point towards single source submarine fan development, a clear focus of sediment supply through the slope (effectively bypassing slope deposition; e.g., Pulham, 1993), whereas the previous elements indicate multiple sources. The predominance of slumped units in the middle part of the Giant Foresets Formation and lack of associated channel systems also points towards a line-source feeder system.

5.6.4 Structural features

(i) Volcanic massifs

The offshore northern Taranaki Basin is characterised by numerous submarine volcanic complexes, dikes and sills, reflecting a prolonged history of volcanism (Mohakatino Volcanic Arc). At least ten individual volcanic complexes have been mapped in the northern part of the basin, six of these within the Northern Graben (Thrasher et al., 2002). On seismic reflection profiles, both volcanic complexes and smaller intrusions are characterised by lower-amplitude, bright, chaotic and discontinuous reflectors. Individual volcanic edifices are further discerned by their clearly mounded profile (Fig. 5.4) and onlapping reflector configuration. Acoustic scattering caused by velocity contrasts between lava flows or volcanic breccias, and underlying sedimentary deposits (Stagpoole, 1997), result in a region of poorly coherent reflectors beneath larger cones. Often it is difficult to trace and correlate sedimentary horizons through these zones. Away from volcanic centres, volcaniclastic sediments (Mohakatino Formation) tend to retain their bright reflector characteristic, and are relatively continuous. The seismic patterns observed in the study area are consistent with Herzer (1995), who noted that buried volcanic massifs in offshore Northland have a characteristic suite of reflection configurations, consisting of (a) a massif with steeply sloping sides; (b) an apron that slopes away from the cone in a smooth, concave upwards curve; and (c) an almost flat ring plain.

Volcanic edifices display either gently peaked or flat topped, steep-sided morphologies, and range in size from small domes, to large complex massifs. The flat-topped nature of some of the larger edifices suggests that part of the structure possibly extended above mean sea level prior to burial, during which time they were bevelled by erosive mechanisms, or that magma composition was such that building of typical 'cones' was not possible. Stagpoole (1997) suggests that in fact the remnants of the largest volcanoes formed an archipelago of islands, or shallow reefs, off the contemporaneous western coast of mainland New Zealand. These later became submerged as down-faulting west of the Turi Fault Zone continued.

Most volcanism had ceased in northern Taranaki Basin by the Late Miocene, and latest Miocene sediments are clearly observed lapping on to those edifices mapped on the seismic grid. Isopach maps (Fig. 5.8a) demonstrate that Pliocene sediment formed only a thin veneer over many of the larger edifices, and that in fact parts of the largest complex were emergent throughout much of the Opoitian (Fig. 5.21). Many massifs continued to exert a physical presence into the Pleistocene, as evidenced in Fig. 5.14b,c, creating topographic highs over which thinner sequences were deposited. It was only in the latter part of the Nukumaruan, after deposition of large volumes of sediment and continued subsidence in the Northern Graben, that volcanic massifs no longer influenced sediment distribution. It is possible that the emergent nature of some of these complexes during the Late Miocene-Early Pliocene contributed to the paucity of sediment of this age in regions to the west (e.g., in the vicinity of Ariki-1, Tangaroa-1), by acting to deflect sediment away from the deeper parts of the basin.

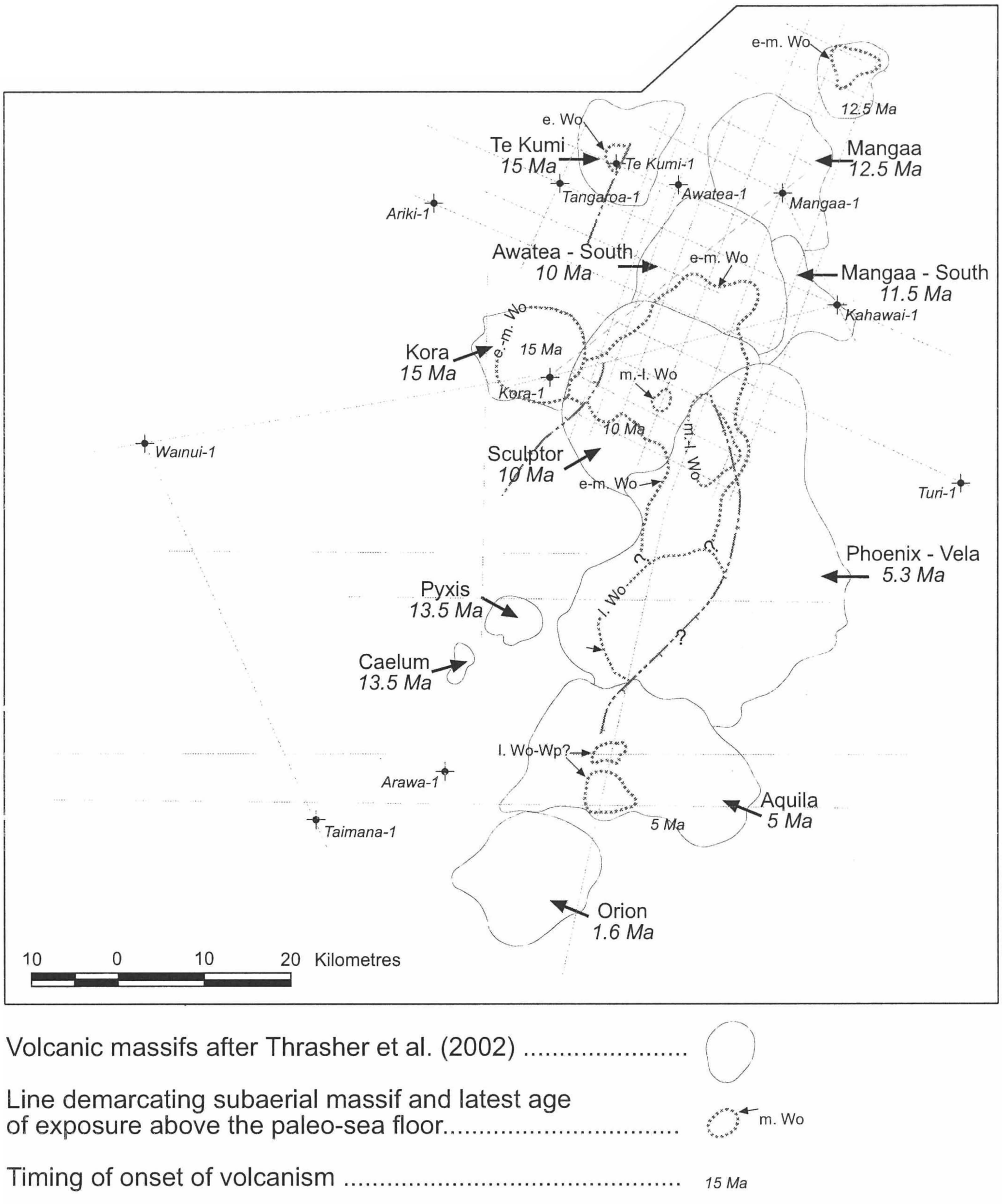


Fig. 5.21: Location of volcanic massifs and timing of onset of volcanism after Thrasher et al. (2002). Parts of some volcanic massifs were upstanding above the paleo-sea floor through to the latest Opoitian (and possibly Early Waipipian), and most continued to create topographic highs into the Nukumaruan.

(ii) Faults

Mid to Late Miocene and Pliocene normal faulting in northern Taranaki Basin occurred primarily as a result of initiation of an extensional tectonic regime. Faults strike predominantly northeast along the Northern Graben, and dip to the west. Since the Pliocene, most fault activity has centered on the diffuse Turi Fault Zone (Stagpoole, 1997), migrating southward with time. Faults in this zone progressively step-up older Miocene sediment to the east, across the Manganui Platform. Across this platform, the Miocene-Pliocene unconformity merges with a regional Pleistocene unconformity. Few faults are identified to the west of the northeast-striking Cape Egmont Fault Zone, and only rarely do faults display offset of seabed or near seabed sediments, indicating that they are presently active (e.g., Nodder, 1993).

Vertical offset on faults is indicated by abrupt termination or rapid change in elevation of a particular reflector or reflectors, bringing rocks of different character into contact, and thus causing sound waves to be reflected (Davis and Reynolds, 1996). Offset ranges in the order of a few tens of msecs (barely resolvable on seismic reflection profiles) to greater than 300 msecs on the larger faults within the Turi and Cape Egmont Fault zones (e.g., Kahawai Fault, line P95-158). It is difficult to determine any strike-slip motion. Most faults within the study area offset older Miocene and underlying sediments, although several of the large faults and fault systems (e.g., Kahawai and Te Kumi Faults, Cape Egmont Fault and Turi Fault Zone) continued to displace sediment into the Nukumaruan.

The reflector characteristics and stratigraphic position and age of seismic units that are affected by faulting are used to interpret the timing of movement on particular faults. Where the thickness of a particular seismic unit is the same on both the up-thrown and down-thrown side of the fault, movement on that fault is interpreted as being post-depositional. Syn-sedimentary growth faults, on the other hand, are delineated by a change in thickness, with the sedimentary unit on the downthrown side of the fault observed to thicken significantly in the direction of the fault. For example, on Kahawai Fault, thickening and a dragging-down effect associated with seismic units A9 to A12 (line P95-158), indicates that faulting was synchronous with deposition of these Opoitian to Waipipian units. Higher in the succession, offset of Late Nukumaruan units indicates that a later phase of offset must have occurred. Further along strike (to the south) the displacement on this fault decreases from 300 msecs (maximum throw) to less than 100 msecs. The few wells in this region (Turi-1, Pluto-1, Mokau-1, and Awakino-1) provide only loose post-Miocene biostratigraphic control, but in conjunction with seismic data, demonstrate that faulting migrated southward during the Late Miocene and Pliocene (see also Stagpoole, 1997). Displacement on other major faults in the study area (e.g., Te Kumi Fault and Cape Egmont Fault) was also occurring during the Opoitian to Waipipian; fault offset shown on structure-contour maps (Fig. 5.13) illustrate how influential various faults were throughout the depositional history of the Giant Foresets Formation.

(iii) Significant unconformities

Unconformities are important in that they represent a hiatus during which erosion and/or non-deposition occurred. The most significant unconformity in the study area is the Miocene-Pliocene unconformity (yellow reflector horizon), which is expressed differently across the study area. In the west, missing time is incorporated in the Ariki Formation – uncertainties about how much time is missing, if indeed any within this highly condensed section, has been discussed in Chapter 3. Further to the south (Western Stable Platform in the vicinity of Arawa-1 and Taimana-1) Miocene and Pliocene strata appear conformable, although a recent review of the biostratigraphy of Taimana-1 has identified a significant break in foreset deposition between mid Opoitian and mid Waipipian strata, indicated by a condensed sequence (Scott et al., in prep.). Throughout the study area, Pliocene sediments lie unconformably (nonconformity) over volcanic massifs. To the east, progressively younger sediments unconformably overlie Miocene strata across the Turi Fault Zone. Within the Northern Graben, the unconformity (paraconformity) is consistent with the yellow reflector horizon, delineated by a bold, laterally continuous reflector.

While the most extensive condensation has occurred over the northern part of the Western Stable Platform (i.e., Ariki Formation), the identification of two condensed horizons in the Northern Graben (upper and lower marly units at Awatea-1 and Mangaa-1) is significant. As discussed in section 5.4.3(i), these units cannot be directly correlated to the Ariki Formation, although they occur in the interval of time covered by the Ariki Formation. This suggests a link between the two regions in terms of the processes that created the condensation.

Figure 5.22 proposes a model for the evolution of the Ariki Formation and the time equivalent horizons identified in the Northern Graben. The basic premise behind this model relates back to sediment starvation caused by the Tangahoe pulldown in Wanganui Basin, as discussed in previous chapters. Sediment starvation appears to be the over-riding factor in the initial development of the Ariki Formation, beginning as early as the Late Tongaporutuan on the Western Stable Platform, and by the mid to Late Kapitean in the Northern Graben. However, commencement of voluminous and sandy siliciclastic sedimentation in the mid Pliocene (Mangaa Formation, lower sands) indicates that this starvation was short lived in the Northern Graben. It is debatable whether this sediment influx was caused by a change (drop) in relative sea level, or whether it was caused by erosion of Whangamomona Group sediments to the east of the study area in King Country Basin.

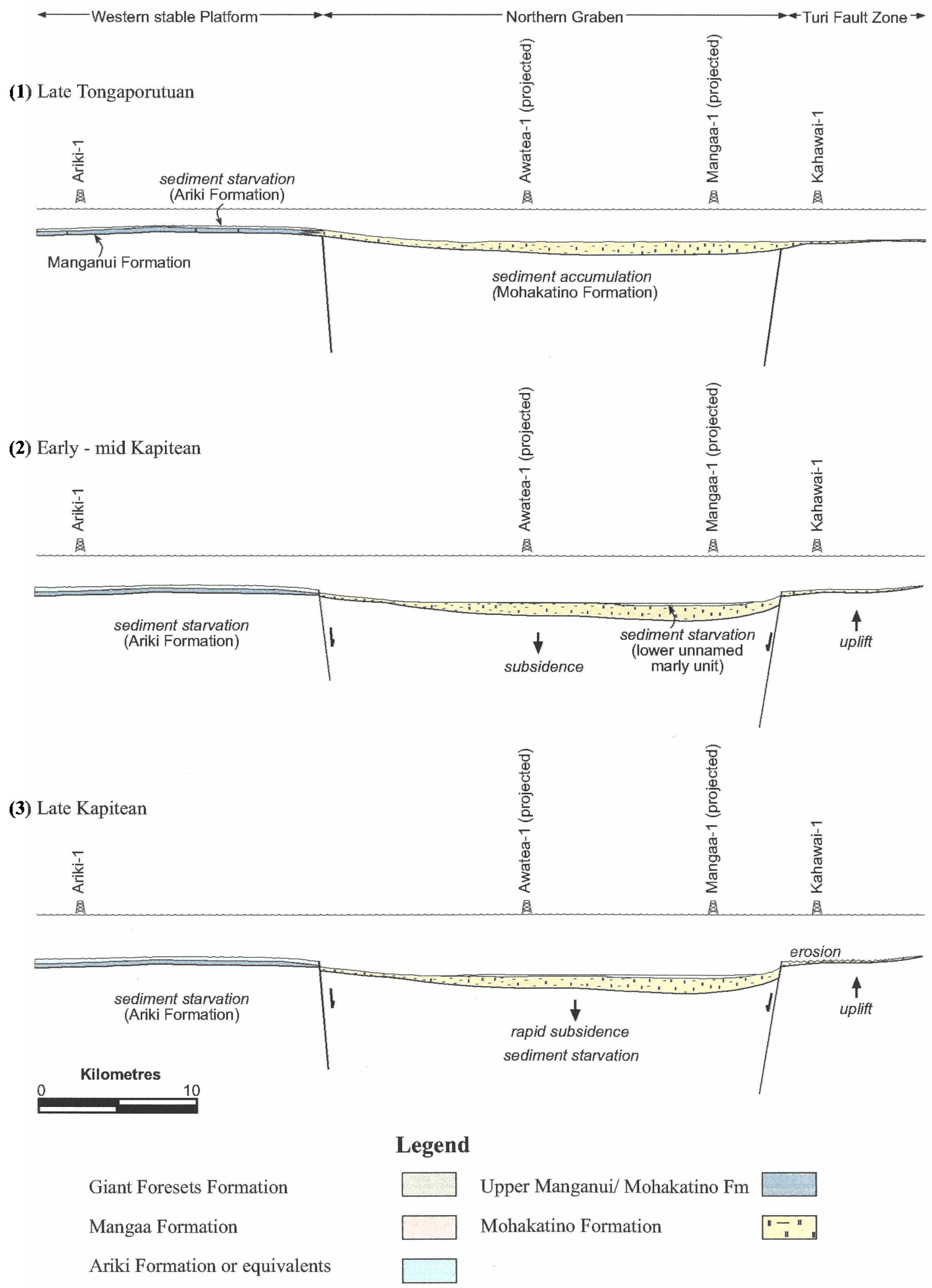


Fig. 5.22: Conceptual model for development of the Ariki Formation and time equivalent condensed horizons, northern Taranaki Basin (continued over page). Not to scale. Thicknesses of condensed units exaggerated. Line of cross-section from Ariki-1 to Kahawai-1.

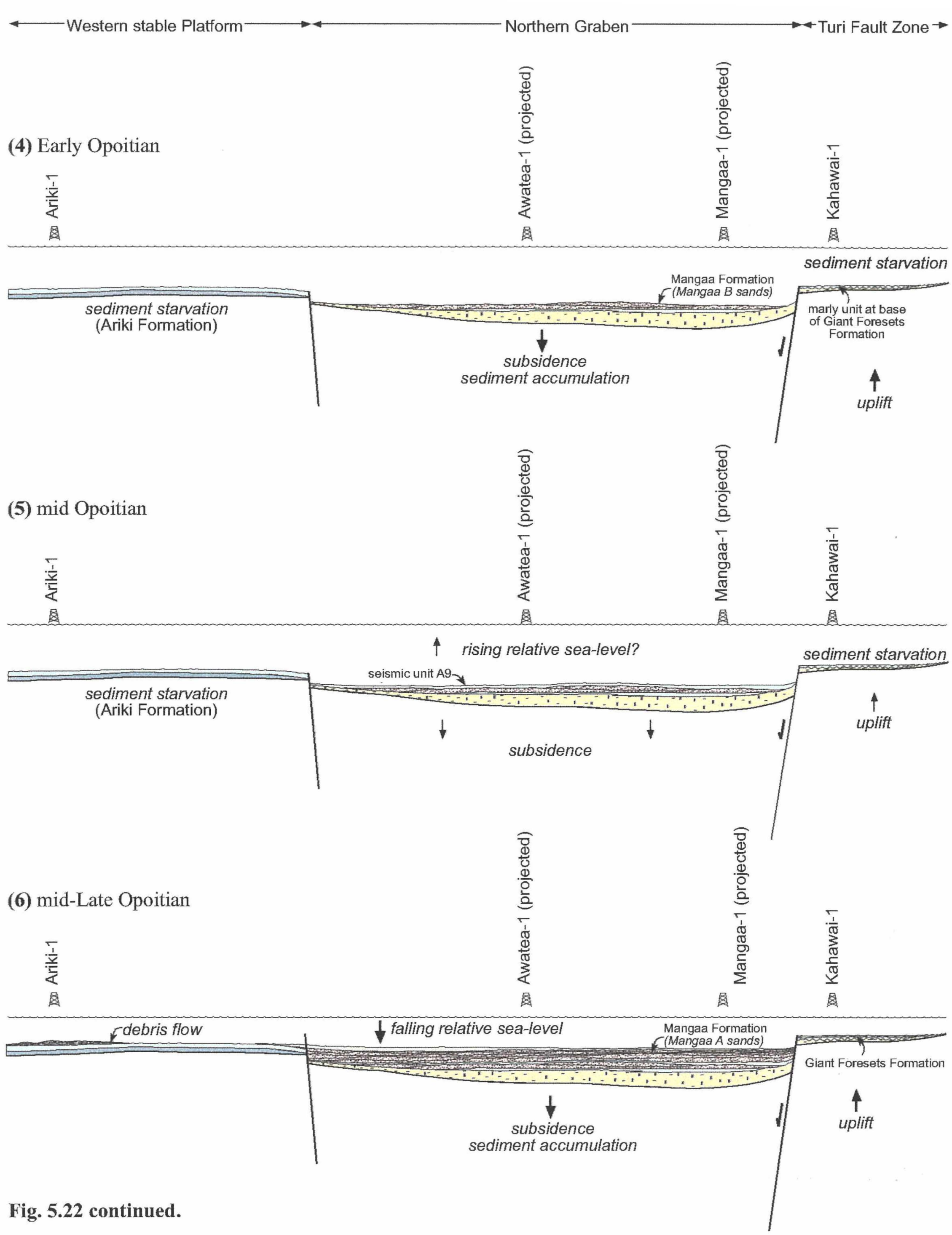


Fig. 5.22 continued.

The presence of a second distinct condensed interval in the Northern Graben (upper marly unit (seismic unit A9) within the Mangaa Formation) suggests that relative sea level change, possibly related to a period of more rapid subsidence, resulted in an apparent flooding event during the Opoitian. Alternatively, the sediment source may have been temporarily shut off during the latter part of the Opoitian, resulting in a second period of sediment starvation. Biostratigraphic resolution is not high enough to indicate the period of time encompassed in this later condensed interval, and depths were such that any relative change in water depth would have to be in the order of hundreds of metres to register changes in planktic and benthic faunal associations. Local factors such as the subtle paleohigh in the vicinity of Ariki-1 relative to the Northern Graben, and uplift on Kahawai Fault further exacerbated sediment starvation, particularly on the Western Stable Platform. It was not until relative sea level began to fall as a result of the encroaching foreset front that sediment began to be deposited in more distal areas.

Undoubtedly, smaller unconformities or diastems (e.g., Nummedal et al., 1993) occur throughout the Plio-Pleistocene succession, associated with movement on certain faults, changing depocentres, or changes in relative sea level (see section 5.8). The condensed interval identified in the Taimana-1 well section is interpreted to relate to a significant break in foreset deposition, and on seismic reflection profiles correlates to the bold reflector horizon at the top of seismic unit B3 (Manganui Formation-Giant Foresets Formation boundary). Above this seismic horizon, seismic units display a significantly different seismic character (i.e., more distinctly mounded external profile, internal downlap; Fig. 5.10) than below this horizon. This is also the case at Arawa-1, and the thin Waipipian section identified in the well section may also suggest a condensed interval.

5.6.5 Distribution patterns and evidence for sediment pathways

Sediment pathways (Fig. 5.23) through the latest Miocene to Pleistocene have been interpreted based on mapped channels and channel orientation, presence and influence of bathymetric highs created by underlying volcanic massifs, major fault traces and the isopach maps presented in Fig. 5.8(a,b). Channel orientation plays a particularly important role in interpretation of events in northern Taranaki Basin. Channel orientation changes from dominantly northwest - southeast (Late Miocene to Early Pliocene channels of Thrasher and Cahill, 1990), to dominantly north-south by the Pleistocene. This change in orientation reflects the timing of Northern Graben formation and it's effect as a sediment trap during the mid/Late Opoitian to Mangapanian. By the Early Nukumaruan the graben had been infilled and then overtopped, with subsequent channels funnelling sediment to the prograding slope to the northwest.

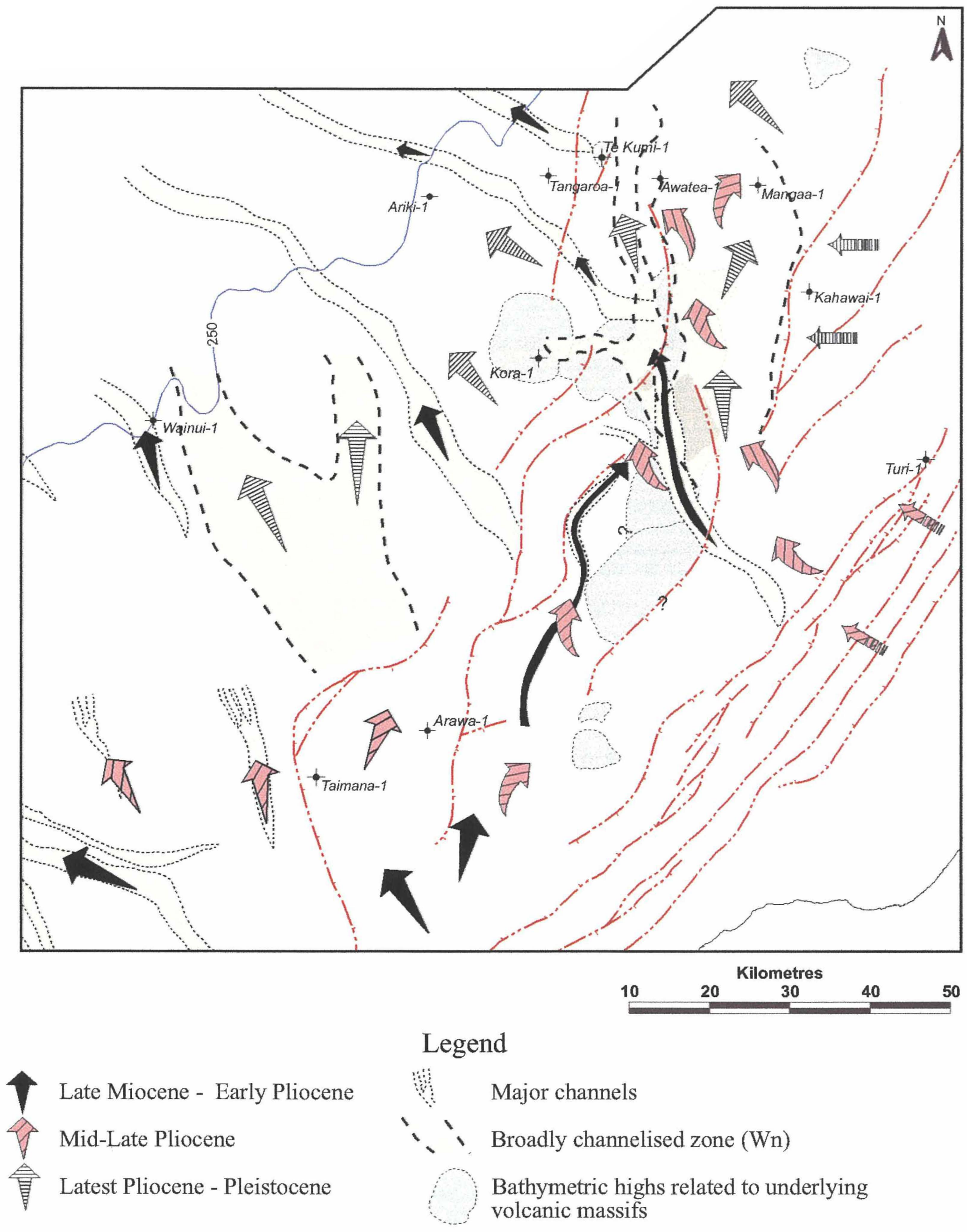


Fig. 5.23: Inferred sediment pathways, Late Miocene to Pleistocene, based on channel distribution (simplified), volcanic massifs (subaerial extent from Fig. 5.21), major fault traces, and isopach maps.

The supply of clastic sediment to the northern Taranaki basin is mainly inferred from offlapping clinoform directions to have been sourced from the south, and is intimately linked to erosion of the emerging Southern Alps. Petrographic studies (e.g., Bergman et al., 1990; Smale, 1996) support a South Island source for Giant Foresets Formation sediments, although Forder and Sissons (1992) suggest that earlier Mangaa Formation sediments were also derived from metasedimentary Murihiku Supergroup rocks to the east of Taranaki Fault. Isopach and structure contour maps, as well as Fig. 5.23, support a northern source, and it is suggested here that the South Island provenance signature is predominantly acquired from uplift and erosion of Whangamomona Group sediments (which have a South Island provenance) from King Country Basin to the east.

5.7 Configuration of the continental shelf margin through time

Taranaki Basin is presently considered to be a passive (stable) margin basin, and displays characteristics typical of this margin type, such as a thick sedimentary cover, which reflects long-term structural stability and slow subsidence (Vanney and Stanley, 1983). Shell (in Balley, 1987) defined the progressive outbuilding of clinoforms, and the resultant shelf-slope profile, as developing into a prominent Atlantic-type shelf margin, where successive clinoform reflectors represent the transient position of the shelf-slope profile (King and Thrasher, 1996).

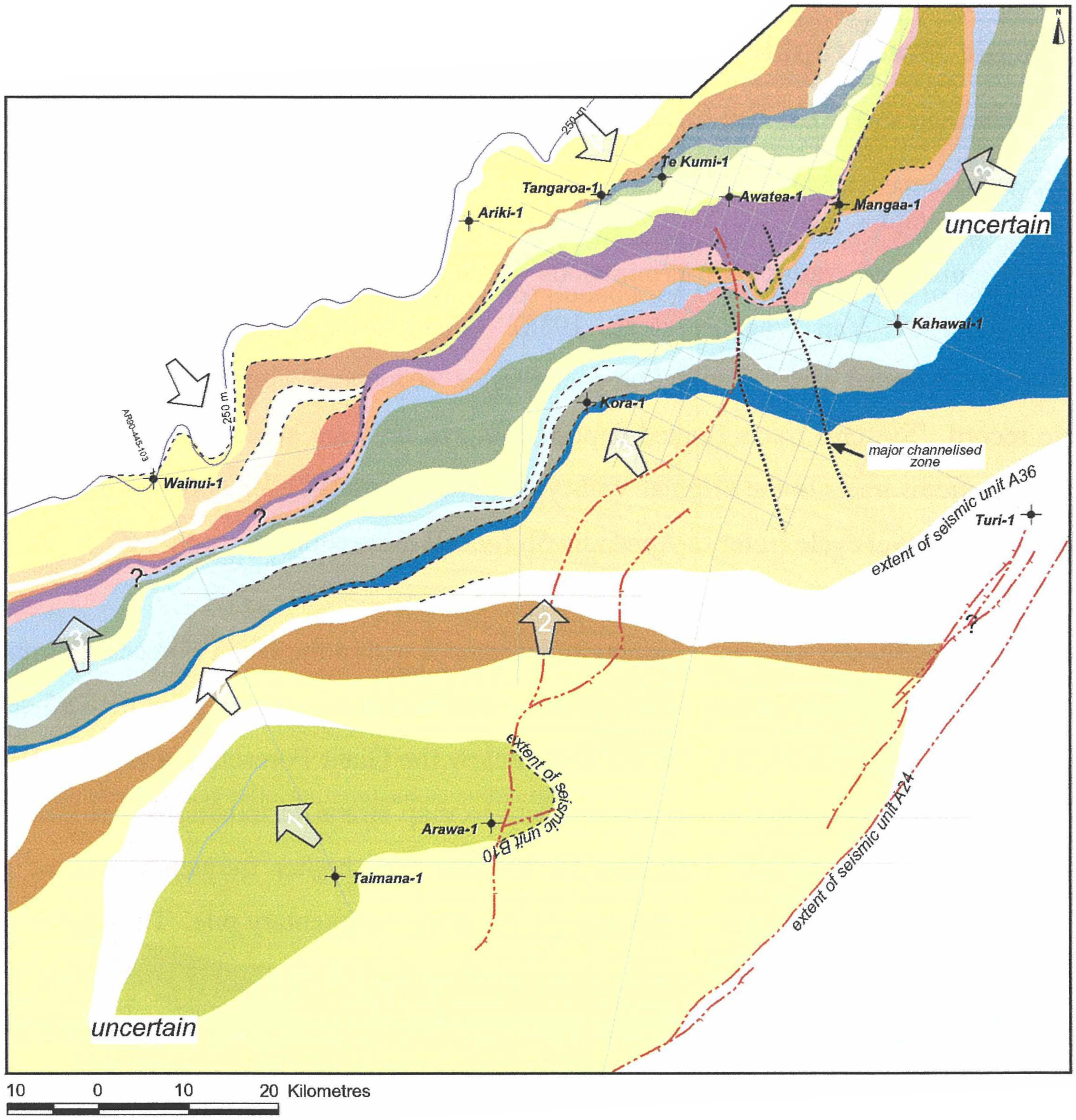
Mapping the approximate position of the paleo-shelf break for each of the seismic units that exhibit (a) progradational clinoform characteristics, and (b) a distinct change in gradient, was an important exercise in terms of understanding the nature of shelf migration through time. The resultant map (Fig. 5.24) demonstrates that while progradation followed a roughly north to south trend, there were several distinct directional changes, or phases, associated with this progradation.

Phase 1 (Waipipian) – Progradation initially occurred in a northwesterly direction as a series of discrete slope-fan lobes (seismic units B4-B7). The concomitant shelf margins for seismic units B4-B7 were south of the study area.

Phase 2 (Late Waipipian-Mangapanian?) – Direction of progradation swung towards the west and north, with deposition occurring along a more linear front rather than as individual fan lobes (seismic units B8-B11, A24-A36).

Phase 3 (Mangapanian to Nukumaruan) – Progradation of the shelf margin continued to swing to the west and north, with the advancing foreset front now aligned in a northeast-southwest orientation (seismic units A37-A55). The shelf margin was beginning to approximate the geomorphology of the modern-day shelf break at this time.

Phase 4 (Late Nukumaruan to Castlecliffian) – The latest phase is characterised by back-stepping (retrogradation) of the shelf margin (seismic units A60-A70), followed by truncation of underlying units landward of their co-eval shelf break (degradational seismic units A73-A73).



Legend

uncertain/obscured

direction of progradation/Phase

Fig. 5.24: Shelf break positions through time, from major seismic units. Note the changing direction, from northwest (Phase 1; Waipipian) to increasingly north (Phase 2; Late Waipipian-Mangapanian?, and Phase 3; Mangapanian-Nukumaruan), and most recently, a period of degradation (Phase 4; Late Nukumaruan-Castlecliffian). See Appendix 5d for seismic unit colour code.

Soenander (1991) also acknowledged these gross trends, identifying a clockwise rotation of the paleo-shelf edge from the southeast to northwest. All of the paleo-shelf breaks that can be mapped in the study area are Waipipian or younger. The contemporary shelf breaks of Opoitian to Early Waipipian seismic units (of which mainly bottomsets facies are identified in the field area) are inferred to be located south of the study area.

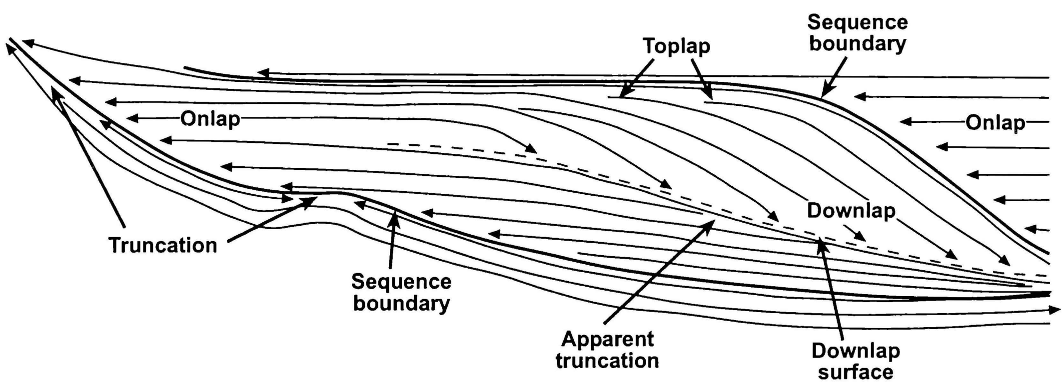
5.8 Seismic stratigraphic interpretation – sequences and controls on stratal architecture within the Giant Foresets Formation

Sequence stratigraphic principles evolved initially from the study of seismic reflection profiles and the recognition of stratigraphic terminations within these types of data (e.g. Mitchum et al., 1977b; Vail, 1987; Van Wagoner et al., 1988). Using these termination configurations, unconformity-bounded seismic reflection packages or 'sequences' could be delineated in the seismic record. Systems tracts (linked depositional systems) are subdivisions of sequences, with three systems tracts in the original Vail-type sequence model, each corresponding to part of a relative sea level cycle (refer to Appendix 5b for full discussion).

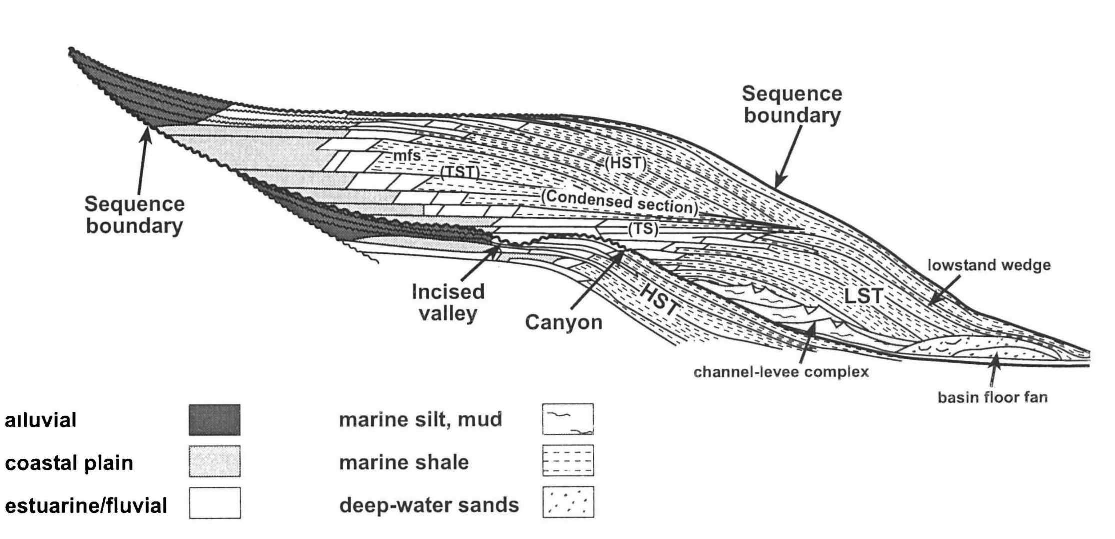
High resolution (100 and 41 k.yr.) mapping of sequences within the Giant Foresets Formation has not been possible in this study. The seismic acquisition parameters (e.g., in the P95-series of seismic lines) were set for leads in the basin below the Giant Foresets Formation. Hence there is inadequate resolution in the seismic data to map stratal terminations and thus map sequences and systems tracts. This may be possible with higher frequency seismic data acquired with parameters focused on the upper 2 km of the sedimentary pile. Thus the seismic reflection profiles have been interpreted and mapped on the basis of highly contrasting gross internal reflection configurations (e.g., Kolla et al., 1984), rather than on the basis of seismic terminations on key reflection surfaces. Nevertheless, a number of the seismic units mapped correspond to seismic sequences as defined in Mitchum et al. (1977b), and thus can be discussed in terms of their sequence architecture and cyclicity. Given these constraints in defining seismic *units* versus seismic *sequences*, only one seismic reflection profile has been interpreted in a sequence stratigraphic context (P95-103). It is therefore important to note that this study is by no means complete, and much work needs to be undertaken to further define the sequence architecture, the total number of sequences that can be identified within the study area, and the resolution of those sequences.

5.8.1 Sequence boundaries – defining seismic units versus seismic sequences

Sequence boundaries form in response to a number of variables, including relative and eustatic sea level fluctuations, tectonism, climate, and sediment supply (Vail, 1987). They represent time-stratigraphic boundaries and thus provide chronostratigraphically significant surfaces that can be used for correlating sedimentary units. On seismic reflection profiles, sequence boundaries are normally recognised by their relationship to reflector terminations (discordance), specifically, onlap, toplap, downlap, and erosional truncation (Fig. 5.25). However, in the study area it is often very difficult to determine if onlap exists because of the frequently disrupted internal reflector configuration characterising many seismic units, and also because onlap is simply not present. While the significance of this is discussed later, these difficulties mean that sequences, and sequence boundaries, are often not able to be confidently resolved.



(a) Sequence-defining reflector termination configurations



(b) Depositional systems and systems tracts

Fig. 5.25: **(a)** Classical seismic sequence-defining criteria, and **(b)** depositional systems model. Adapted from Vail (1987).

Recent studies (e.g., Steckler et al., 1993; Kolla et al., 1995) have suggested that the timing of formation of the sequence boundary can vary from the time of maximum rate of sea level fall to the time of the lowest position of eustatic sea level; in particular, faster tectonic subsidence rates and higher rates of sediment supply may cause the timing of the sequence boundary to be delayed (Galloway, 1989). However, in a seismic sequence stratigraphic context, the easiest reflector horizon to define and follow is the one defined by onlap in the slope area. In this case, the boundary is traced below that part of the slope-basin profile that is termed the basin floor fan. Thus, at least on seismic reflection profiles, the lower sequence boundary appears to occur beneath the basin floor fan (cf. wireline sequences), as consistently interpreted by Exxon workers.

The sequence boundary on the continental shelf in the Vail model is interpreted to occur at the top of the HST. In the study area HST deposits, where they can be identified in sequences, are present on the shelf; however, invariably a thin siltstone that accumulated during a highstand will be present on the slope and basin floor, and may correspond to a condensed horizon that, on seismic data, is boldly defined. Where present, this clinoform reflector can be bright, displaying moderate to high amplitude reflectivity and continuity from basin floor through to shelf. Beggs (1990) and King and Thrasher (1996) interpret these clinoforms as reflecting horizons of partial lithification of the paleo-sea floor, representing flooding surfaces and/or condensed sections formed during depositional hiatuses associated with relative high sea levels. Thus the clinoforms approximate sequence boundaries.

The delineation of sequence boundaries is further assisted by the recognition of several features and architectural elements attributed to falling or low relative sea level conditions, such as the occurrence of downlapping reflectors and the occurrence of incision by submarine canyons and channels, interpreted to occur during the falling limb of a relative sea level cycle. Sequence boundaries within the Giant Foresets Formation are thus delineated on several criteria:

1. the occurrence of bold, high amplitude reflectors that are relatively continuous from shelf to basin floor;
2. onlap of strata onto the lower sequence boundary;
3. erosional truncation of strata by the sequence boundary, in conjunction with criteria 1 and/or 2 above; and
4. the occurrence of slumped packages, as these often overlie sequence boundaries.

5.8.2 Sequence architecture and systems tracts

5.8.2(i) The classic model

In the classic Vail model (Fig. 5.25) each seismically defined depositional sequence has a predictable set of systems tracts and bounding stratal surfaces and systems tracts (defined in Appendix 5b), including (in ascending order) a sequence boundary, a LST, a transgressive surface, a TST, a maximum flooding surface, and a HST (Vail, 1987). Early ideas of a sequence (e.g., Mitchum et al., 1977b; Vail et al., 1977a,b; Vail, 1987; Posamentier and Vail, 1988) were very model driven, but more recent studies (e.g., Galloway, 1989; Posamentier and Allen, 1993) have recognised that the variability in processes and depositional products of deep marine settings means that sequence stratigraphy must be viewed as a tool or approach rather than a rigid template. Particularly important in any sequence stratigraphic model is local physiography, tectonics, sea level fluctuations and the type and rate of sediment supply (Posamentier and James, 1993; Emery and Myers, 1996).

5.8.2(ii) Typical sequences in the Giant Foresets Formation

Few depositional systems and stratal architectures conform entirely to the block diagrams presented in Posamentier et al. (1988), with the Giant Foresets Formation being no exception. Rather than rigorously adhering to the approach exemplified in Fig. 5.25, this study has taken more of a stratigraphic approach to identifying components of a sequence and the timing of their accumulation in terms of a sector of the relative sea level curve. This approach is based on the recognition of architectural elements as discussed in section 5.6, the depositional systems that these elements represent, and thus their appropriate ordering within a sequence stratigraphic framework.

Figure 5.26 presents a sequence stratigraphic interpretation of seismic reflection profile P95-103. This seismic line illustrates several seismic unit boundaries that correspond with sequence boundaries as defined by the criteria outlined in section 5.8.1, and a break down of internal features using the architectural elements approach. Sequence boundaries are clearly delineated by bright, high amplitude reflectors, occasional onlap, erosional truncation, and where basin floor fans are present, by downlap. Most sequences are characterised by concordant reflectors from shelf to slope, although evidence of planar down-cutting is associated with seismic unit A54 and seismic unit A70. Concordance from slope to shelf facies is evidence of a lack of subaerial exposure and erosion.

The sequences illustrated on Fig. 5.26 are typical of late Nukumaruan strata, and comprise several readily identifiable architectural elements, the most common of which are basin floor (lower slope) fans and channel-levee complexes. Often, the sequence boundary underlies Type-2a basin floor fans, which sometimes display bi-directional downlap on to the lower sequence boundary, or Type-2b fans which grade conformably into hemipelagic basinal facies. Basin floor and lower slope fan facies are concordant with channel-levee complexes of the middle slope. These latter architectural elements are relatively extensive, forming stacked complexes of channels and levees separated by subparallel reflectors of overbank deposits.

Legend for Fig. 5.26 (over page)

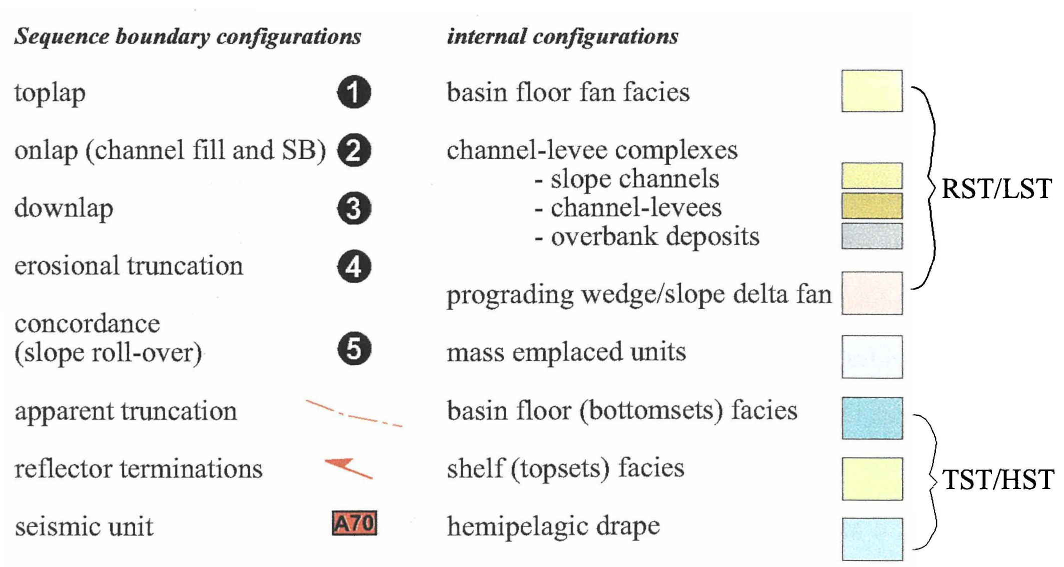
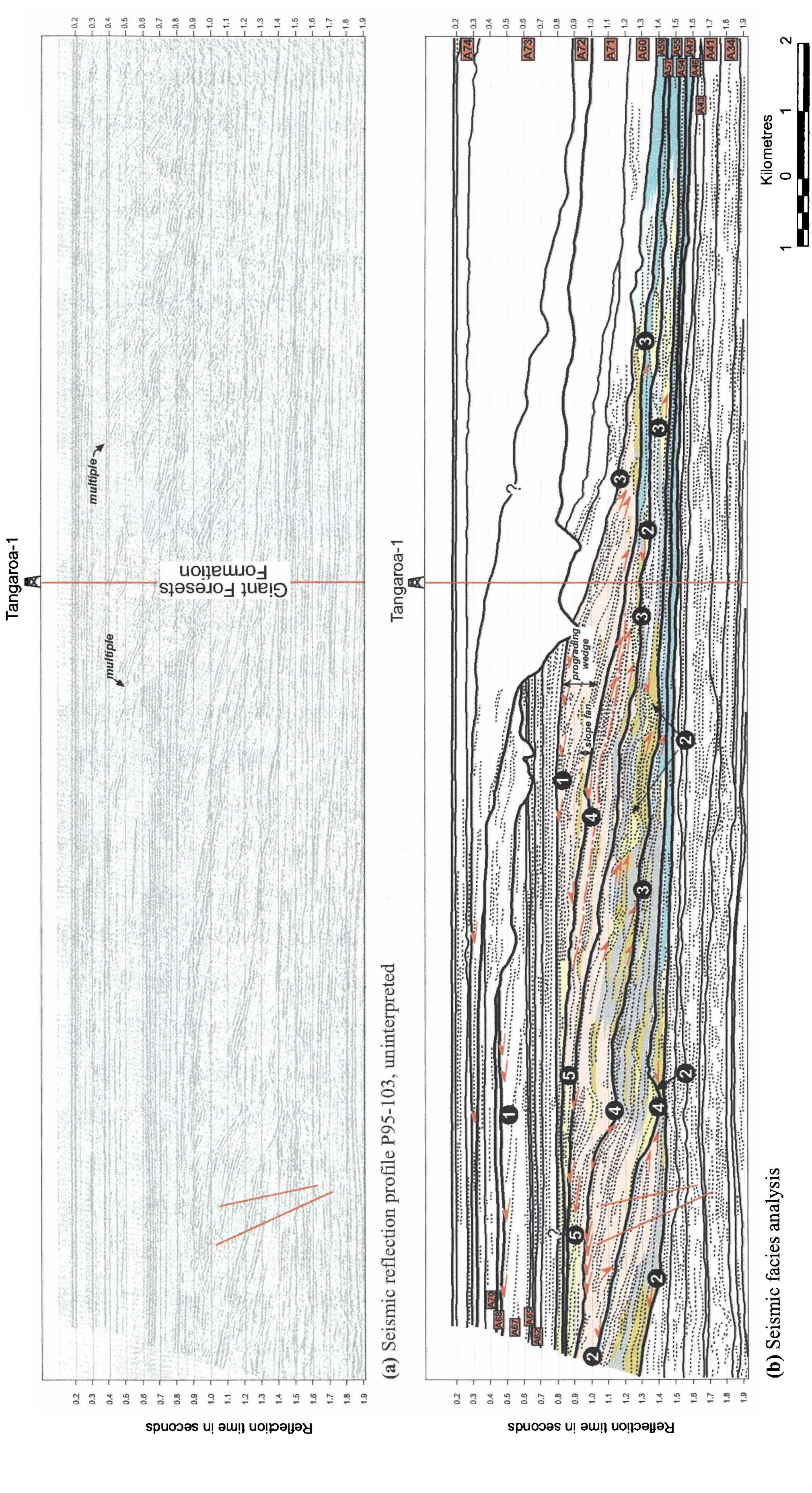


Fig. 5.26 (facing page): Seismic sequence interpretation, Giant Foresets Formation. **(a)** Uninterpreted line (P95-103); **(b)** sequence interpretation of line P95-103 (based on recognition of specific architectural elements). Sequence boundaries are delineated by bold lines - these can only be confidently placed where a complete or almost complete shelf-slope profile is present. Basin floor fan facies (lowstand fans) are more typical of mixed mud-sand/mud-rich provenances, with low mounds characterised by relatively low-amplitude to reflection free internal configurations. Low relief gull-wing geometries of channel-levee complexes coupled with acoustically transparent patterns characterise the slopes. Each sequence displays (occasional) onlap, toplap, and downlap, as well as erosional truncation.



(a) Seismic reflection profile P95-103, uninterpreted

(b) Seismic facies analysis

Studies of modern deep-marine clastic systems have suggested that fan deposition can occur during any stage of the relative sea level cycle (e.g., Galloway, 1989), but the volumes of clastic sediment delivered to the basin are likely to be highest during periods of falling or low sea level (Emery and Myers, 1996). During this interval, remobilisation of sediment at the shelf edge delivers detritus to deeper water in the form of slides, slumps, debris flows and turbidity currents, by-passing the slope and being deposited on the basin floor. During the stillstand and early part of sea level rise, sedimentation occurs in the foreset region (channel-levee complexes). While sea level fall is stressed in sequence stratigraphic models, it is recognised that sediment caliber and flux, as well as local basin physiography, have a greater impact on fan evolution and stratigraphic expression than published models allow (Posamentier and Allen, 1993; Emery and Myers, 1996; Eschard, 2001). The dominance and volume of muddy sediment delivered to the northern Taranaki Basin is thus reflected in the stratigraphic expression of the lowstand depositional units (architectural elements) that accumulated basinward of the shelf break, with a lack of large sand-dominated fan lobes, few deeply incised channels, and numerous channel-levee complexes.

Sequences within the Giant Foresets Formation display an abundance of architectural elements interpreted to have been dominantly formed during lowering and low relative sea level (i.e., fans and channel-levee complexes). Several sequences include a thin set of reflectors downlapping on to fans and channel-levee complexes that may be interpreted as lowstand wedge deposits; this diverges from the typical Vail-type model, which shows a very thick lowstand wedge relative to basin floor and channel-levee components. Depositional architectures similar to those within the Giant Foresets Formation are observed on the New Jersey Continental margin, where Miocene-Pliocene sequences lack lowstand wedge deposits (Metzger et al., 2000), and often display a marked lack of deeply incised canyons or channels (Fulthorpe and Austin, 1998; Metzger et al., 2000). The absence of these units is attributed to a lack of exposure of the shelf break during relative sea level lowstands. Several theories have been postulated to explain this observation, including long term rise in sea level minimizing the effect of high frequency sea level falls and/or because a very high sediment supply caused the shelf to continue to build seaward during sea level lowstands (rather than being incised), with by-passing of lowstand sediment into deeper water (resulting in a lack of lowstand wedge).

The stratigraphic extent of the transgressive systems tracts, which in sequence stratigraphic models are characterised by backstepping parasequences, are difficult to map seismically in the topsets (shelf facies) of the Giant Foresets Formation. With most sequences, the TST is unable to be separated from the HST, and may be represented by a single reflector. On the shelf, the TST/HST may be up to 50 msecs thick, implying rapid sedimentation during rising relative sea level and early relative sea level fall (Lawrence, 1993). The thickness of the TST/HST on the shelf is a function of topset accommodation volume, which in turn is a function of the magnitude of (relative) sea level rise multiplied by the topset area (Milton and Bertram, 1995). The balance between sediment supply rate and the rate of creation of topset accommodation volume is thus thought to control cycles of progradation and retrogradation – in brief, the wide northern Taranaki Basin shelf, being widest during maximum sea level transgression, created enough space to accommodate all transgressive and highstand sediment so that none, or very little, reached the slope and basin floor.

Basinward of the shelf break, the HST tends to merge with the sequence boundary, but may occasionally be represented by the presence of hemipelagic drape on the paleoslope (e.g., seismic unit A54). However, drape configurations also occur within sequences, for example, as the result of lobe switching on fans (e.g., Crews et al., 2000). With seismic unit A54, the bold and continuous reflectors suggest that in this case, the drape is the result of a high relative sea level.

Although not every seismic line will encounter every systems tract, comparison with the classic Vail-type seismic sequence stratigraphic model illustrates the asymmetrical nature of Giant Foresets Formation sequences versus the typical clastic-dominated sequence architecture commonly discussed in the literature (discussed further in section 5.8.6). This is further highlighted by Fig. 5.27, a chronostratigraphic reconstruction of seismic reflection profile P95-103. This series of panels, particularly panel c, helps to further refine the position of sequence boundaries. It also illustrates clearly the progradational nature of the foresets facies, and the degree of erosion associated with the degradational foreset facies.

The sequences that have been identified in P95-103 (using Figs. 5.26 and 5.27) have been correlated across a northwest-southeast transect, on to which the broad lithology has been added (Fig. 5.28). This transect illustrates that many of the interpreted sequence boundaries occur immediately below coarse lithological units, and that within a single sequence, topset/ uppermost slope facies are generally coarser than mid slope facies.

Legend for Fig. 5.27

Unconformity - approximate position of sequence boundary ....

Topsets (shelf facies) ....

Foresets (clinoforms) ....

Slope channel fill ....

Bottomsets ....

Basin floor fans ....

Marine condensation ....

Erosion due to truncation/channel incision ....

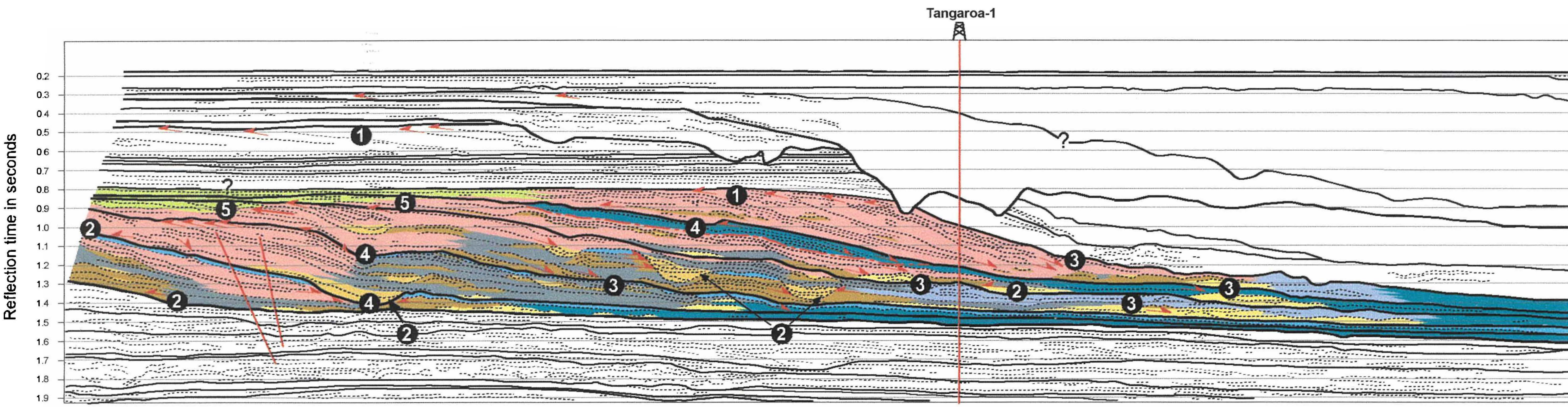
Fig. 5.27 (following page): Chronostratigraphic reconstruction (Wheeler diagram) of seismic reflection profile P95-103. Construction is based on the methodology outlined in Emery and Myers (1996, their figure 5.1). This figure highlights a number of features, including the progradational nature of clinoforms, the highly erosional nature of the degradational foresets, and the degree of marine condensation, indicating most of the sedimentary thickness is accumulated during lowering and low sea level conditions.

5.8.3 Controls on progradation

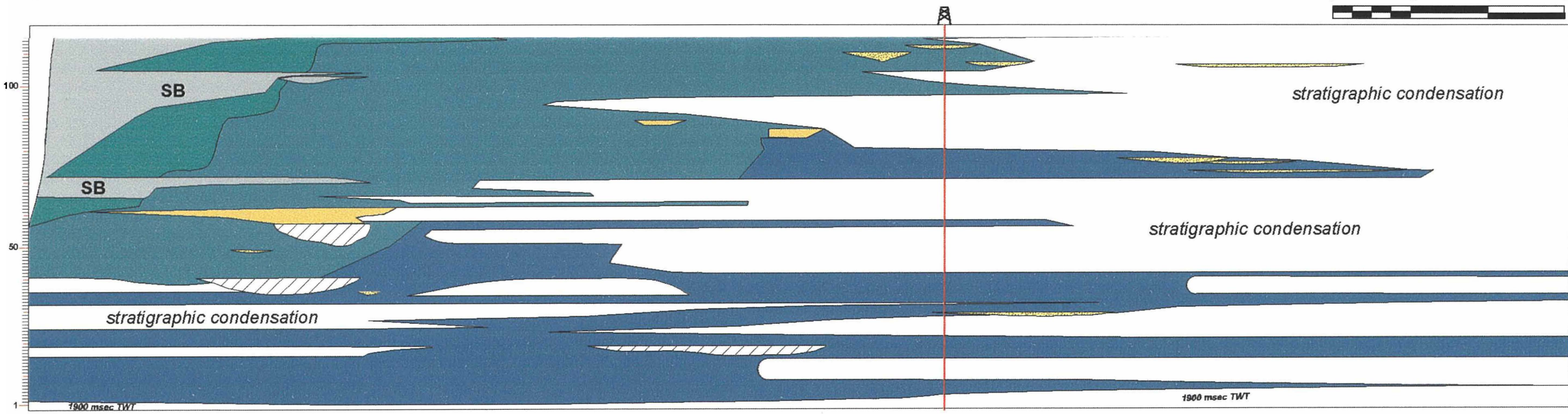
A sequence architecture develops in response to changes in the rates of relative sea level rise and fall, reflecting the addition of the related subsidence and eustatic curves (Jervey, 1988; Posamentier et al., 1991). The concept of accommodation is fundamental to sequence stratigraphy, and can be quantified by the following equation (from Emery and Myers, 1996):

$$\Delta\text{accommodation} = \Delta\text{eustasy} + \Delta\text{subsidence} + \Delta\text{compaction}$$

While eustasy and subsidence rate control the amount of accommodation, sediment supply fills the accommodation available and controls water depth. The relative sea level curve is hence a measure of the rate of change in accommodation. The depositional architecture is expressed as a function of the accommodation created (Fig. 5.29). Vail (1987) considers the fundamental control on depositional sequences to be short-term eustatic changes of sea level superimposed on long-term tectonic changes.



(a) Seismic sequences and facies analysis (from Fig. 5.25; see figure for legend).



(d) Chronostratigraphic interpretation of line P95-103. Legend on facing page.

Fig. 5.27: Legend on facing page.

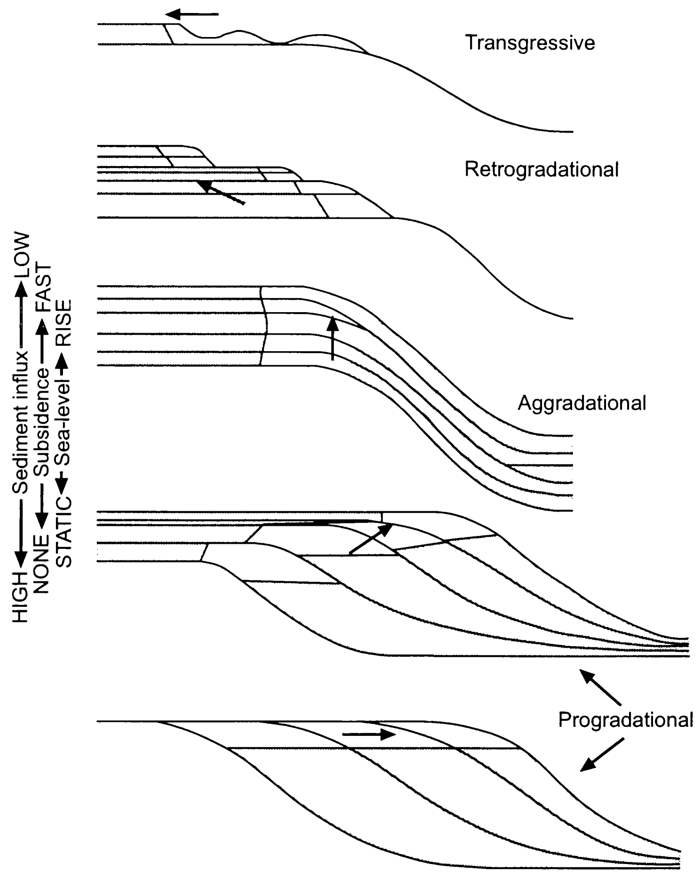


Fig. 5.29: Clinoform stacking patterns as a function of accommodation volume and sediment supply (after Galloway, 1989; Emery and Myers, 1996).

5.8.3(i) The role of tectonism in the Northern Graben

The influence local tectonism exerted on the study area, especially in the Northern Graben region, is particularly evident on seismic reflection profiles and associated depositional maps. While essentially a passive margin basin, renewed subsidence associated with extensional tectonism during the Late Miocene and Early Pliocene created new accommodation space in the Northern Graben, in which sediment was trapped and rapidly accumulated, as illustrated by isopach maps Fig. 5.8a,b.

Burial history plots of selected wells (Fig. 5.30; discussed more fully in Chapter 7) illustrate several key features. These include rapid subsidence from the Pliocene onwards (less so at Wainui-1 than within the graben), as well as clearly displaying the rapid shallowing following progradation of the slope into the study area (paleo-sea bottom curve). Much of the subsidence can be attributed to sediment loading. Subsidence rates reduce during the Nukumaruan, leading to the accumulation of topsets.

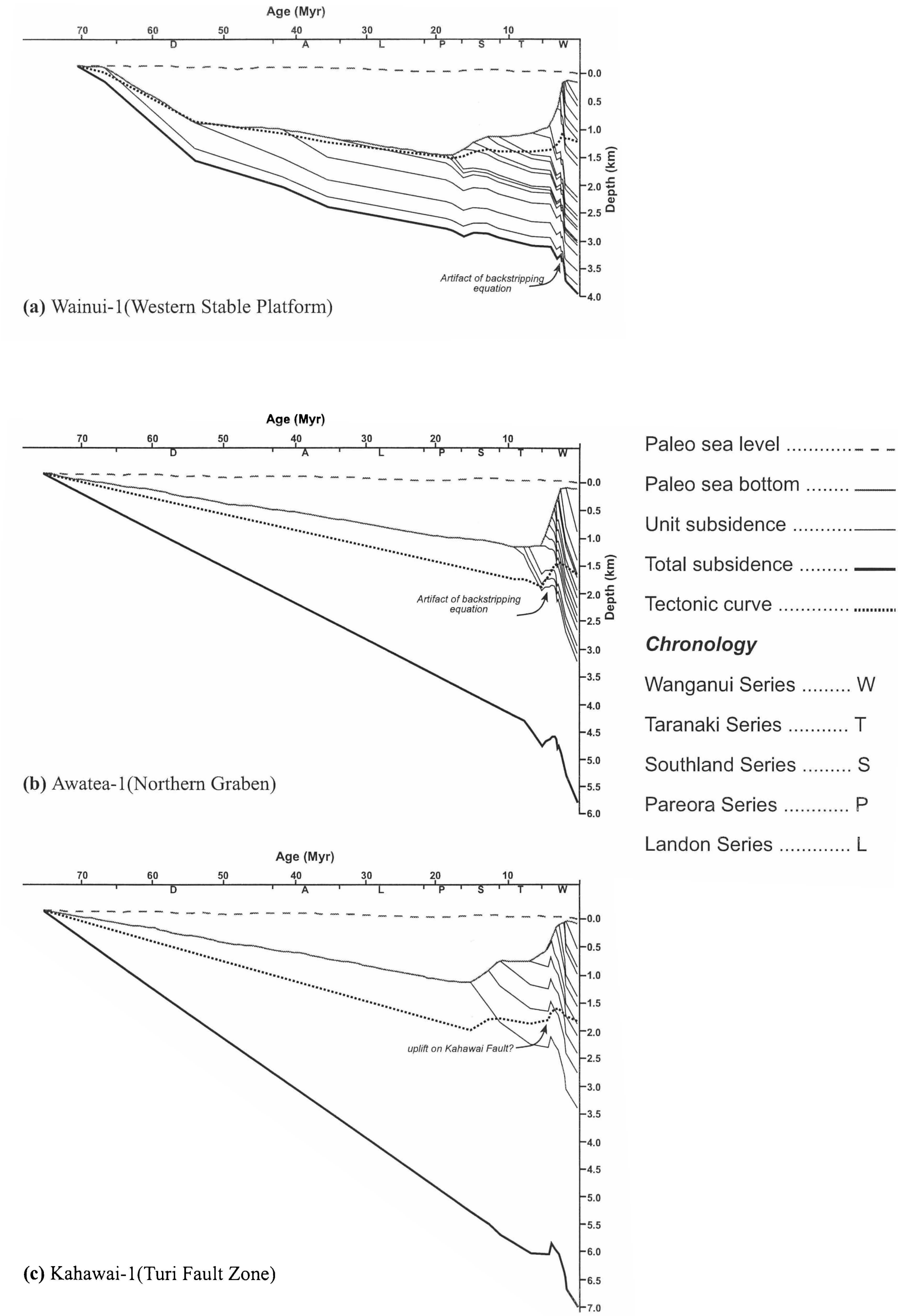


Fig. 5.30: Geohistory curves of selected wells from (a) Western Stable Platform (Wainui-1), (b) Northern Graben (Awatea-1), and (c) Turi Fault Zone (Kahawai-1). See Chapter 7 and Appendix 7 for backstripping information.

Although tectonic subsidence in the graben and uplift on the Turi Fault Zone continues to the present day, differential fault-related subsidence played an increasingly less important role towards the end of the Pliocene and into the Pleistocene. Sediment supply began to outstrip subsidence rates due to increasing sediment volumes being sourced from the Southern Alps (Tippett and Kamp, 1993), and possibly also from an increase in uplift and denudation of the King Country region of the North Island. Seismic reflection profile P95-158 does however show evidence of a later period of subsidence in the northern part of the graben, associated with movement on Kahawai Fault. This movement occurred during the Late Nukumaruan, and resulted in compensation-style deposition of seismic units A61-A68 (Enclosure 3, P95-158).

5.8.3(ii) Eustatic sea level change

It is difficult to quantify the amount of accommodation space created and lost during eustatic sea level rises and falls, and indeed the magnitude of sea level change itself, in a deepwater setting. While a rise or fall in sea level of 100 m or more on the mid and inner shelf may be plainly evident in the dislocation of facies belts, this same magnitude has limited impact on sedimentation when water depths are of the order of hundreds of metres or more. This is further compounded by the occurrence of a continually subsiding basin, the rate of which may approach the rate of sea level fall. Backstripping (Chapter 7) and forward modelling (e.g., Kamp and Naish, 1998) are some of the methods of analysis that can be used to estimate changes in relative sea level through time and predict sequence architecture, but unique solutions require the magnitude and frequency of eustatic sea level change to be known.

A number of workers have attempted to construct global sea level curves (Vail et al., 1977b; Haq et al., 1987, 1988). An alternative model of eustatic sea level change, at least for the Pliocene and Pleistocene, is provided by astronomically tuned oxygen isotope ($\delta^{18}O$) curves. Figure 5.31 presents an $\delta^{18}O$ curve for the Late Miocene through to Recent obtained for DSDP Site 593 (Challenger Plateau – see Fig. 6.21, Chapter 6 for location; Cooke, 2002). This smoothed curve records repeated periods of warming and cooling, with an overall increase in ice volume, and therefore lower sea level, through the Pliocene and Pleistocene, consistent with the dominantly regressive Giant Foresets system. While the magnitude of change in sea level at DSDP Site 593 cannot be calculated, using an isotope curve obtained from Huon Peninsula, Papua New Guinea, Chappell and Shackleton (1986), and Chappell et al. (1996) have calibrated sea level during the last glacial maximum (MOI2) as 120 m lower than present day sea level –

the most extreme glacial of the Quaternary. This is similar to a value obtained by Proctor and Carter (1989) who state that at the peak of the last glaciation, sea level was approximately 113 m lower than present, turning Cook Strait into a northwest-southeast aligned embayment.

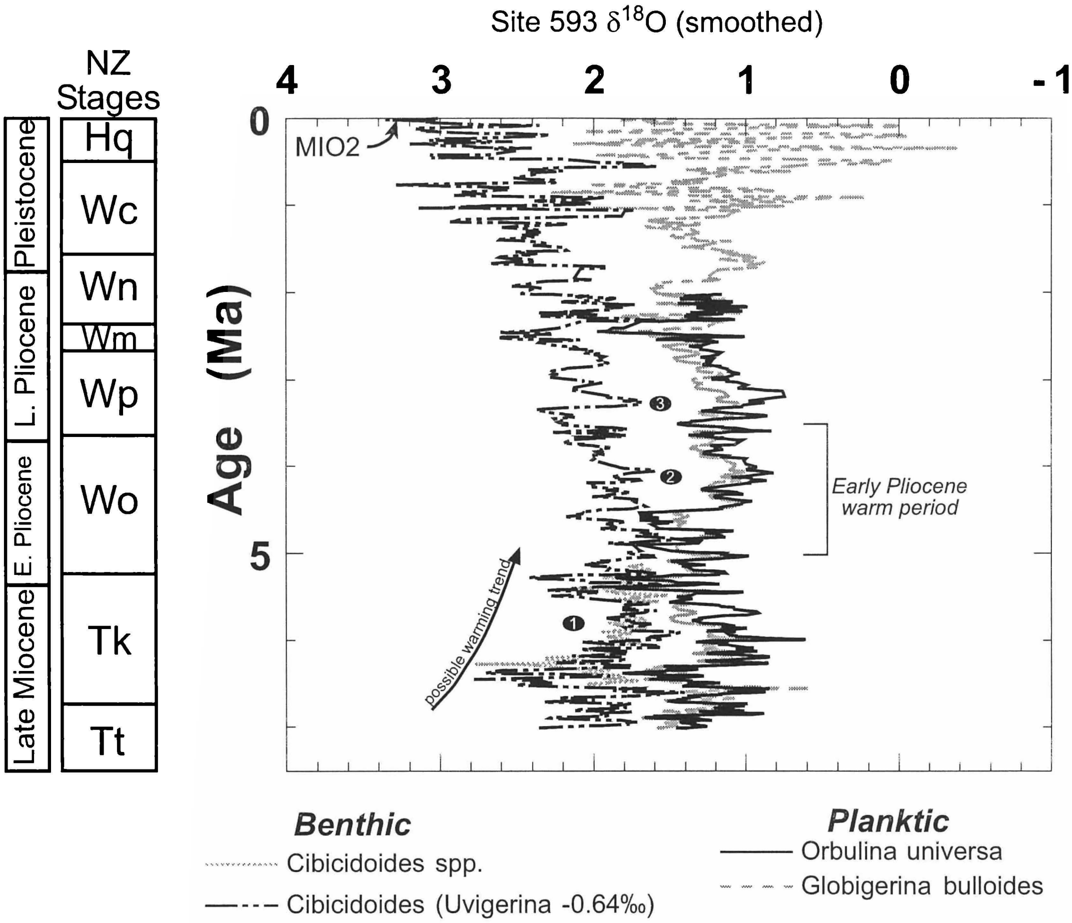


Fig. 5.31: Oxygen isotope curve for DSDP Site 593 (adapted from Cooke, 2002). Note the general warming trend during the Late Miocene to Early Pliocene, and increasing magnitude of cooling events in the Quaternary. Possible flooding events discussed in text are marked (1, 2, and 3). MIO2, Marine Oxygen Isotope stage 2. Tt, Tongaporutuan Stage; Tk, Kapitean Stage; Wo, Opoitian Stage; Wp, Waipipian Stage; Wm, Mangapanian Stage; Wn, Nukumaruan Stage; Wc, Castlecliffian Stage; Hq, Haweran Stage.

While sediment starvation is postulated as the main control on accumulation of the Ariki Formation, and at least the lower marly unit immediately underlying the Mangaa Formation, the warming trend (lower ice volumes) observed through the Kapitean Stage and into the Opoitian Stage (marked by a ‘1’ on Fig. 5.31) may have exaggerated the degree of condensation. Correlation between seismic unit A9 (upper marly unit within the Mangaa Formation) and a warming trend in the Mid to Late Opoitian (marked by a ‘2’ on Fig. 5.31) suggest that the upper marly unit may be linked to a flooding event. A third warming trend in the Waipipian (marked by a ‘3’) may help explain the abrupt termination of coarse siliciclastic sedimentation in the

Northern Graben. However, these links are only very tentative, as no $\delta^{18}O$ data have been obtained from any of the well sites.

The greater magnitude eustatic sea level lowstands implied by the most enriched $\delta^{18}O$ values (i.e., larger ice volumes) during the Pleistocene may have impacted significantly on the nature of the Giant Foresets Formation – lower lowstand sea levels, coupled with a decreasing subsidence rate in the Northern Graben would mean that, during lowstands, the shoreline would be much closer to the shelf break than during earlier stages. This would result in more sediment being delivered to the shelf edge, leading to increased slope instability, slumping, and degradation (e.g., seismic units A72 and A73).

5.8.3(iii) Sediment supply

Sediment supply and volume has exerted a direct influence on the stratal patterns observed in seismic reflections profiles. and contributed directly to the large area reclaimed from bathyal to shelf depths by deposition of the Giant Foresets Formation (Beggs, 1990). Denudation rates calculated for the Southern Alps during the Plio-Pleistocene (~0.5-2.5 mm/yr.) indicate that a crustal section up to 18 km thick has been eroded during the last ~ 8 Ma (Kamp et al., 1989; Kamp and Tippet, 1993; Tippet and Kamp, 1993). A large amount of erosion is also suggested for proximal North Island source areas, with a few hundred metres to 2.5 km of denudation associated with up-doming of the King Country Basin estimated from the Pliocene to present day. These estimates are preliminary only, and are not well constrained (Kamp et al., in prep.), but together with South Island-derived detritus, emphasise the volume of sediment that was available for distribution.

The present-day littoral drift system moves 2000-4500 $m^3/y \times 10^3$ of sand and gravel north along the west coast of the South Island, but does not cross over to the North Island (Williams, 1985). Deeper currents similarly remove large amounts of sediment from the west coast of the South Island, but these are presently funnelled into Cook Strait, *not* Taranaki Basin, while smaller volumes of sediment (100-500 $m^3/y \times 10^3$) are distributed to the north and south of Taranaki Peninsula by the North Island's west coast littoral drift system (Williams, 1985; De Lange, 2002, pers. comm.).

While this depicts the present-day highstand situation, modelling by Proctor and Carter (1989) indicate that during Pleistocene glacial maxima, the emergence of a land bridge between the North and South Islands (portrayed as a broad coastal plain with a western shoreline connecting Taranaki Peninsula with Cape Farewell) would have severely altered coastal dynamics, disrupting tidal flows and changing sediment distribution patterns. Consequently, during lowering/low sea levels, sediment derived from the South Island would have been diverted to the west and north. This is consistent with the implication that much of the outbuilding of the continental margin occurred during low relative sea level conditions, and concurs with the conclusions of Bosario (1981), who demonstrated that widening of the continental shelf (south Taranaki Basin) predominantly occurred during Pleistocene glaciations, with an increased sediment flux derived from uplift and glacial erosion carried northwards by the D'Urville Current. Furthermore, it implies that during the Pleistocene, eustatically-controlled sea level change played a larger role in the magnitude of relative sea level change than did subsidence. It also suggests that during Pleistocene highstands, sediment derived from North Island axial ranges may have been a locally important sediment source.

The physiographic setting during much of the Pliocene was somewhat different than that during most of the Pleistocene. Originally part of a broad shelf during the Early Pliocene, rapid subsidence of the Wanganui Basin during the Early Pliocene (Tangahoe Pulldown; Kamp et al., 2002) created a depocentre that entrapped sediment derived from South Island uplift and erosion. The terrigenous sediment that did reach the northern parts of the study area (i.e., the Mangaa Formation) at this time is inferred in this study to have been sourced from the more proximal North Island. Through the Waipipian to Early Nukumaruan, the Wanganui Basin depocentre was rapidly infilled (Tangahoe Mudstone and younger units). Subsequently, progradation that had begun earlier in the south rapidly advanced into northern Taranaki Basin.

While changes in relative sea level control the amount of accommodation, sediment flux along a basin margin and the proximity of sediment entry points to the basin affect the rate at which that accommodation space is filled, directly influencing both the lateral and vertical distribution of facies belts (Jervey, 1988; Wehr, 1993; Emery and Myers, 1996). Importantly, progradational geometries will only occur when the basin margin is supplied with a greater volume of sediment than available topset accommodation (Milton and Bertram, 1995; see Fig. 5.29). Although both sediment flux and eustatic sea level changes produce the seismic character and architecture observed within the Giant Foresets Formation, concordance between slope and shelf sediments and updip thinning of sequences indicates that an increased rate of sediment flux over eustatic sea level fall exerted a dominant control on the stratal architecture of foresetted units (e.g., Lawrence, 1993). Increasing sediment flux during falling sea level conditions is thus the key to the asymmetry of sequences interpreted on line P95-103, much as it is within shelf sequences identified in the Wanganui Basin (e.g., Naish and Kamp, 1997b).

Seismic units A61 to A68, although truncated landward of their concomitant shelf break, give the impression of overall aggradation. Aggradational geometries occur when sediment supply and the rate of creation of topset accommodation volume are roughly balanced, resulting in vertical stacking and a static shelf break (Milton and Bertram, 1995; Fig 5.29). As previously mentioned, seismic units A61-A68 display large-scale compensational-style bedding, associated with infilling of a depositional low created by renewed subsidence in the Northern Graben during the Late Nukumaruan. This subsidence would have created additional accommodation space and a local rise in relative sea level.

The position of seismic units A69-A71 relative to underlying units (see Enclosure 3, line P95-103 and P95-158), and the planar truncation of seismic units A66-A68 landward of their associated shelf break suggests a brief period of retrogradation prior to accumulation of seismic units A69-A71. This retrogradation suggests that for a short interval sediment supply was less than the rate of generation of topset accommodation, resulting in erosion of the underlying units as the sedimentary wedge began to prograded basinward again. Progradation accelerated with the deposition of units A72 and A73 (degradational foresets); the huge thickness and volume of sediment associated with these units suggests they are a result of a gross imbalance between accommodation space and sediment supply, possibly linked to the more extreme glacial maximums of the middle to Late Pleistocene. The coarser lithologies associated with the degradational foresets facies represent a change in depositional regime from the mud-rich system of underlying units, to a mixed mud- sand-rich system, and probably reflects an increasing sedimentary contribution from the North Island.

5.8.4 Relative sea level and the shelf break – how low did sea level fall?

The stratal patterns associated with sequence boundaries, and the internal reflectors themselves, may constrain the magnitude of relative sea level fall. While King and Thrasher (1996) considered that periodic falls in relative sea level were of a magnitude great enough to lower sea level below the preceding shelf break, this study concludes that, for the most part, this does not appear to have been the case in northern Taranaki Basin. Several lines of evidence support this conclusion:

1. The majority of seismic units/sequences display internal concordance from slope to shelf, indicating continuity of shelf and slope sedimentation (roll-over).
2. Topsets (shelf facies) are rarely truncated, except within the uppermost part of the profile (e.g., planar down-cutting of seismic unit A70 and A71). Truncation of units underlying seismic unit A73 may be related to the approach of the contemporary shoreline to the shelf break, but may also be the result of mechanical failure due to sediment texture and over-supply.
3. Based on present day water depths (approx. 200-250 m at the shelf break), and taking into consideration continuing subsidence (Fig. 5.30), relative sea level fall would need to be in excess of 200 m to expose the shelf break. Although a fall of 120 m is recorded during the Quaternary (MOI2; Chappell et al., 1996), $\delta^{18}O$ values indicate that prior to this, sea level change was not of this magnitude.
4. Facies belts do not display the drastic facies dislocation one would expect with such a decrease in water depth to the shelf edge (i.e., inner shelf or shoreface sediments overlying slope sediments).
5. Absence of deeply incised canyons or channels, other than that associated with seismic unit A55, indicate that sea level was rarely low enough to initiate major incision.

These conclusions are consistent with foraminiferal analyses (Chapter 6), which indicate that, at least until the Late Nukumaruan to Castlecliffian, relative sea level fluctuations were not great enough to expose the shelf break.

5.8.5 Regressive versus transgressive sequences

Classic sequence stratigraphic models (e.g., Vail, 1987; Van Wagoner et al., 1988) such as illustrated in Fig. 5.25 display similar thicknesses of lowstand and highstand successions. While it is recognised that the highstand component is over exaggerated, most models imply that sedimentation during relative high sea levels accounts for much of the sedimentary thickness of a sequence. Furthermore, systems tracts models hold only if variations of sediment supply are small and eustatic oscillations dominate the system. If supply is allowed to vary significantly, the systems tracts no longer reflect the changes in accommodation but reflect the dynamic equilibrium between changes in accommodation *and* supply (Schlager, 1993).

Sequences within the Giant Foresets Formation diverge from 'classic' models in that they are asymmetric, with sedimentation occurring mainly during regression and lowstands. The slope facies contain very thin deposits relating to rising and high sea level conditions. This observation is analogous to the asymmetrical shelf sequences observed in Wanganui Basin, in which thick regressive systems tracts have been recognised (Naish and Kamp, 1997b). This departure from the classic model is interpreted to be predominantly a function of marked changes in rate of sediment supply and type, modulated by climatic and relative sea level change.

5.8.6 Sequences or multisequences? – A question of resolution

The ability to identify the fundamental sequence signal, and to determine their order of cyclicity, is dependant on the resolution of seismic reflection profiles and dating methods. For example, detailed outcrop studies of the Mount Messenger Formation section exposed along the northern Taranaki coastline have resulted in the identification of numerous sequences of various frequencies, including 3rd, 4th, and 5th order cycles with periodicities of 1 m.y., 400 k.y., and 100 k.y., respectively (e.g., King et al., 1994; Hansen, 1996; Browne et al., 2000). In the Wanganui Basin, 6th-order (41 k.y.) sequences have been identified from integrated subsurface wireline log and outcrop studies (e.g., Naish and Kamp, 1997a,b; McIntyre, 2001; Kamp et al., in prep.). These previous studies have the advantage of direct observation and correlation from outcrop to subcrop. As outlined in Chapter 4, the situation is somewhat more complex in the offshore study region because of (a) an obvious lack of outcrop, (b) predominantly muddy lithologies, and (c) paucity of sampling through shelfal (topsets) facies. This makes it difficult to define sequences based on wireline log and lithofacies patterns and characteristics.

Seismic reflection profiles provide a way of defining sequences because each sequence is delineated by specific reflection configurations, and although within the Giant Foresets Formation these are not exactly the 'classic' sequence-defining criteria, nevertheless several sequences can still be identified. However, unlike wireline log or outcrop studies, the resolution of individual reflectors (i.e., the ability to tell that more than one feature is contributing to an observed effect; Sheriff 1985) means that only lower frequency cycles can be delineated. This is because a seismic wavelet displayed in travel time may represent 30 to 50 m or more of geologic strata (Davis, 1992). In the Wanganui Basin, higher frequency (6^{th}-order) eustatically-driven sequences occur in the order of 10-100 m through the shelfal Matemateaonga Formation and parts of the Rangitikei Supergroup. These points suggests that it is virtually impossible to recognise 6^{th}-order (41 k.y. duration) cyclicity in industry-acquired seismic reflection profiles from northern Taranaki Basin. It may be possible to recognise 5^{th}-order (100 k.y. duration) cycles if they were thick enough, but a rough estimate of the number of seismic units that could also be sequences, versus the time-frame through which they occur suggests that the order of cyclicity that is displayed by the Giant Foresets Formation clinoforms is in fact closer to 4^{th}-order. This concurs with King and Thrasher (1996) who suggest that individual clinoform sequences have a similar depositional cyclicity to Mount Messenger Formation lowstand sequences described along the northern Taranaki coast section.

Correlative basin studies (i.e., in Wanganui Basin) have determined that higher frequency eustatically-controlled cycles are a feature of the Pliocene and Pleistocene stages (dominated by 41 k.y. periodicity prior to 0.9 Ma, and 100 k.y. periodicity subsequently; Kamp et al., 2002). Thus, in the context of the seismic sequences that can be recognised in the Giant Foresets Formation, it must be assumed that each 'sequence' most likely comprises many higher frequency cycles, the signals of which cannot be confidently or consistently isolated via the interpretative techniques utilised in this study. This was highlighted in Chapter 4 (section 4.6.3), which discussed the fact that, although possible 5^{th}-order cyclicity could be identified in some wells (e.g., Tangaroa-1, Mangaa-1), these cycles were not evident in *all* wells, and therefore could not be correlated on the basis of wireline signatures alone.

Sequence boundaries are evidence of depositional hiatus, corresponding to an interval of geologic time that is not represented by strata. Where this interval is significant (e.g., a million years or so), the surface is an unconformity (Mitchum et al., 1977b). In most instances, sequence boundaries within the study area realistically would not represent more than a few tens of thousands of years, and are more likely to be time-rich horizons of condensation, rather than periods of non-deposition. Unfortunately, biostratigraphic and lithologic description of well cuttings are generally not at a high enough resolution to pick up either minor unconformities or condensed sections. Wireline log signatures do appear to be able to delineate some of the more obvious, higher amplitude and brighter bounding reflectors, for example, the top of seismic unit

A54. This boundary corresponds to low GR, higher velocities, and a kick in SP and resistivity values, and was also identified as a sequence boundary on the basis of wireline signature (e.g., Tangaroa-1, Chapter 4). Other seismic units display similar wireline responses, but again, these are not consistent throughout the succession.

5.9 Summary: seismic characteristics of the late Neogene sedimentary succession and inferences about the evolution of the continental margin

External and internal characteristics of seismic units, the distribution and thickness of these units, and any cyclicity that can be determined provide valuable insight into the evolution of the late Neogene sedimentary succession in northern Taranaki Basin. Late Miocene sedimentation involved the accumulation of the Manganui Formation, characterised by moderate to low amplitude internal reflectors that are often subparallel to wavy and discontinuous. These deposits are often intercalated with bright and laterally continuous reflectors of the Mohakatino Formation, which are observed to onlap Miocene volcanic massifs.

Late Miocene to Early Pliocene sediment starvation in the study area is expressed by the accumulation of a condensed marl over the northern part of the Western Stable Platform (Ariki Formation). Seismic mapping of the Ariki Formation has revealed its limited extent, although two other condensed units have been identified in the Northern Graben. Also within the graben, Pliocene sediments unconformably onlap Miocene volcanic massifs. Miocene strata are truncated across the Turi Fault Zone.

The Mangaa Formation is of middle to Late Opoitian age and accumulated within the Northern Graben soon after its formation. The sediments were probably sourced from erosion of Whangamomona Group strata in western North Island. The sandstone beds of the Mangaa Formation created a series of low relief, linear basin floor fan mounds, characterised by bright, high amplitude and laterally continuous external and internal seismic reflectors. Characteristic thickening of seismic units A10 and A14-A19 towards Kahawai Fault indicate that subsidence was ongoing during deposition of the Mangaa Formation.

Rapid progradation of the continental margin from south to north was initially expressed in the study area as a series of discrete Late Opoitian to Waipipian slope fan lobes (seismic units B4-B7). These were deposited in the vicinity of Arawa-1 and Taimana-1, and are characterised by a mounded external geometry, smooth parallel to subparallel internal reflectors, and bi-directional downlap on to a basal reflector (top of seismic unit B3). The seismic characteristics and lobate aerial extent of these fan mounds indicate a point-source feeder system. Isopach and shelf break maps indicate that progradation initially occurred in a northward direction, but became progressively oriented along a more linear southwest-northeast front.

Increasing sediment flux into the study area was marked by accelerated progradation of the foreset front across the study area during the Late Waipipian to Nukumaruan, and is also indicated by a change of seismic character of the progradational seismic units. Rather than involving the continual accumulation of slope fan mounds, the succession became dominated by the deposition of classic sigmoidal progradational clinoforms.

Seismic units B1, A19, A24-A30, and A34-A40 are characterised by highly disorganised to chaotic internal reflectors, intercalated with reflection-free zones, low to moderate, occasionally discontinuous bounding reflectors, general lack of distinctive termination patterns, and frequent absence of any recognisable reflector configurations that could be interpreted as basin floor fan or channel-levee architectural elements. While the external profile is initially smooth and the slope inclination low, successive units thicken basinward and display increasingly steeper inclinations, with a concurrent decrease in the smoothness of their external profile, and an increase in the disordered nature of internal reflectors. Oversteepening, instability and slope failure at the shelf break became prevalent, with lower boundaries characterised by slump scarps and arcuate shear planes. The seismic character of these seismic units, and general absence of major channels, indicates a mud-rich system, with slumps and debris flows the dominant sediment transport mechanism.

Seismic units A41-A57 and B13-B17 are characterised by continuous, often bold, moderate to high amplitude bounding reflectors, and low to moderate amplitude internal reflectors characteristic of channel-levee complexes and muddy basin floor fans. Turbidity currents are interpreted as the dominant depositional mechanism, although occasionally slumping is still locally important (e.g., mounded toe of seismic unit A55). Numerous meandering and straight channels systems are also associated with these seismic units. The seismic characteristics, presence of multiple channel systems, and linear depositional front suggest a mixed mud/sand-rich depositional system fed by a linear source area.

Shingled clinoform configurations and large-scale compensation style bedding displayed by seismic units A60-A70 indicates progradation of the shelf margin during relative high sea levels and continuing subsidence of the Northern Graben. These units hint at a period of decreased sediment flux and possible retrogradation of the shelf margin; however, lack of evidence of retrogradational backstepping because of planar downcutting by subsequent units (seismic units A71-A73) means that this can only be postulated.

The extreme degradation and vast thickness displayed by seismic units A71-A73 indicate that the paleo-shoreline migrated much closer to the shelf edge during Pleistocene low sea level conditions, and possibly even reached it. While each degradational unit must incorporate a number of cycles of sea level change (nearly 2 m.y. may be represented by these three units), these are difficult to differentiate due to the internal disruption associated with each unit. Detailed provenance studies have never been conducted on Giant Foresets Formation sediments, but it is possible that the change in seismic character relative to the character displayed by older foreset units is a combined result of (a) greater influence of relative sea level change on the shelf break due to regional shallowing; (b) a higher proportion of coarse siliciclastic sediment delivered to the shelf edge; and (c) a change in sediment type and texture due to an increasing contribution from a (?volcaniclastic) North Island source.

5.10 Conclusions

- The Giant Foresets Formation is well imaged on seismic reflection profiles. Seismic units can be divided into topset, foreset (progradational and degradational) and bottomset beds (after Beggs, 1990). While underlying formations are generally characterised by parallel to subparallel reflectors, the Giant Foresets Formation is characterised by a diversity of reflector characteristics, suggesting several depositional mechanisms.
- Progradation into the study area initially occurred as a series of individual fan lobes, probably fed from a single point source. The distribution pattern of sediment associated with the advancing continental margin gradually changed to a more linear relationship, reflecting increasing shelf width, increasing sediment flux, and decreasing influence of local tectonics. Structure-contour, isopach, and shelf break maps demonstrate the overall north-northwest translation of depocentres, and illustrate how the depositional loci changed throughout the Plio-Pleistocene.

- Classic sigmoidal-shaped progradational seismic units of the Giant Foresets Formation typically comprise a number of architectural elements that reflect sediment flux and the dominant lithology. Older (Waipipian-Mangapanian) seismic units are characterised by chaotic, internally disrupted to reflection free configurations consistent with slumping. Progressive oversteepening led to slope failure, while absence of channels and the presence of failure chutes characterise a mud-rich line-source slope apron system with an increasing sediment supply. Younger progradational units are distinguished by numerous small channel-levee systems on the lower and middle slope, and larger meandering channel systems on the shelf and upper slope. Basin floor fans are characterised by low-relief mounds and/or gull-wing geometries of channel-levee systems, or sub-parallel reflectors. These characteristics are more typical of a mixed mud/sand-rich depositional system.
- Mapping of the time-rich Ariki Formation over the northern Western Stable Platform, and identification of two seismically unrelated but age-equivalent condensed intervals within the Northern Graben indicate that sediment starvation, possibly accentuated by several warming trends/flooding events in the Late Miocene to Early Pliocene, impacted substantially on the evolution of the sedimentary succession in the study area. Paleobathymetric relief may have been locally important in extending the duration of sediment starvation.
- The presence of voluminous siliciclastic deposits in the Northern Graben during the mid Opoitian to Waipipian (Mangaa Formation), and the paucity of equivalent aged sediment elsewhere in the study area, indicates that the Northern Graben was an efficient and important depositional sink during this period. Channel orientation and interpreted sediment pathways indicate that the Northern Graben continued to be a focus for sediment deposition during the Late Pliocene and Pleistocene.
- Identification of architectural elements within the Giant Foresets Formation has allowed the preliminary identification of several 4^{th}-order (400 k.y.) sequences. These sequences contain architectural elements that are linked to mechanisms and depositional styles that are more frequently observed with lowering and low relative sea level (RST and LST), such as basin floor (lower slope) fans, and channel-levee complexes. Basinward of the shelf margin, the HST of each sequence is represented by a bright, high amplitude and laterally continuous reflector that rolls over into shelf deposits. Transgressive systems tract components are not recognised basinward of the shelf margin, and are incorporated within the HST landward of the shelf margin.

The HST (and TST?) is time-rich relative to the RST/LST, resulting in a highly asymmetrical sequence architecture.

- The predominant factor controlling the seismic character of the Giant Foresets Formation during the Late Opoitian to Mangapanian appears to be sediment flux and lithology over relative sea level change. However, during the Nukumaruan to Castlecliffian, decreasing accommodation and rapidly shallowing paleobathymetry as a result of accelerated progradation meant that changes in relative sea level increasingly influenced stratal architecture.

Chapter 6: Paleoenvironmental interpretation of the Giant Foresets Formation, northern Taranaki Basin

Chapter 6: Paleoenvironmental interpretation of the Giant Foresets Formation, northern Taranaki Basin

6.1 Introduction

The aim of this chapter is fourfold; (i) to derive a detailed paleobathymetric curve for four well sections within the study area, enabling backstripping and decompaction calculations to be made for palinspastic reconstructions (Chapter 7); (ii) to understand sedimentary and oceanic controls on foraminiferal distribution; (iii) to reconstruct depositional environments to aid in the construction of paleogeographic maps, and (iv) to identify any oscillations in sea level as recorded in the occurrence of sequences.

The four well sections utilised here are the same as those discussed in Chapter 3 (Arawa-1, Ariki-1, Kora-1 and Wainui-1). Mangaa-1, as with Chapter 3, is also used as a control well, primarily because of its particular location (northeastern quadrant of the study area) and because the recent study by Waghorn et al. (1996) has provided relatively detailed information regarding the paleobathymetry and watermass conditions for this site through the Neogene.

6.2 Background information

Several features of foraminifera make them important in paleoenvironmental reconstructions. They are taxonomically numerous, abundant throughout the marine environment, evolved rapidly, are small in size, and able to be recovered from well core and cuttings, and certain species are environmentally sensitive (Boltovskoy and Wright, 1976). Because of their utility in paleodepth reconstructions, much literature is available on relationships between benthic and planktic foraminifera, and their depth distribution in oceans (see e.g., Murray, 1973; Boltovskoy and Wright, 1976; Murray, 1991a,b). In this chapter, a brief introduction to the planktic (6.3.1) and benthic (6.3.2) foraminifera is provided. The reader is referred to the publications given above, and to the various references listed throughout this chapter, for more in-depth discussion on this diverse subject.

Planktic foraminifera live in the water column, and have long been used to determine watermass conditions over a site (Murray, 1976). As such, they have also been used as a proxy for depth, as surface watermasses tend to become more oceanic with increasing distance from land and depth also generally increases away from the shoreline (see section 6.3.1(i)). Benthic foraminifera, on the other hand, inhabit the seafloor, either within the substrate (infaunal) or upon it (epifaunal). Substrate may vary from loose sediment, to firm surfaces such as provided by rocks or shells, and plants such as seaweed.

Benthic foraminifera occur throughout wide ecological ranges, and for many decades their distribution was inferred to be controlled primarily by depth. Studies over the last three decades (e.g., Haq and Boersma, 1978; Berger and Diester-Haass, 1988; Van der Zwaan et al., 1999; Murray, 2001) have shown that their distribution is dependant on the sum of several ecological factors, such as salinity, hydrostatic pressure, temperature, light intensity, current strength, turbidity, nutrient levels, biological competition, pH, O_2 concentration, $CaCO_3$ concentration, and trace elements (Hayward, 1986). Organic flux (food) and oxygen are thought to be particularly important governing factors (Gooday, 1988, 1994; Altenbach, 1988; Berger and Diester-Haas, 1988; Van der Zwaan et al., 1999). Because many species prefer certain ecological niches with particular depth-controlled environmental factors, they are in effect somewhat depth-regulated. It is this element, coupled with the response of certain species to a sudden or dramatic change in one or more of the ecological conditions, which make benthic foraminifera extremely useful in determining paleodepths, ocean circulation patterns, and even processes related to sedimentation, especially within single depositional basins (Murray, 1991a).

Sample preparation

Samples were prepared using the standard method outlined in Appendix 3a. Size of the picked split (>150 µm) for Ariki-1, Kora-1 and Wainui-1 was determined after discussion with M. Crundwell (IGNS) and follows studies such as Scott and Crundwell (1994) and Naish and Kamp (1997b). Although recent studies have argued for the use of the > 63 µm fraction (Schröder et al., 1987; Van der Zwaan et al., 1999) so as to include juvenile specimens and smaller species, a consideration behind picking the >150 µm size fraction was one of time. Arawa-1 had been split and picked (> 125 µm) prior to this study, although species identification for many individuals had yet to be determined, as well as intervening samples picked for planktic-benthic ratios. Where possible, at least 100 benthic individuals were identified and mounted on slides. While many other studies have used upwards of 200 specimens, Hayward et al. (1999) state that 100 benthic specimens are able to provide a sufficiently accurate assessment of dominant faunal composition to identify species associations and to aid in constructing paleoenvironmental maps.

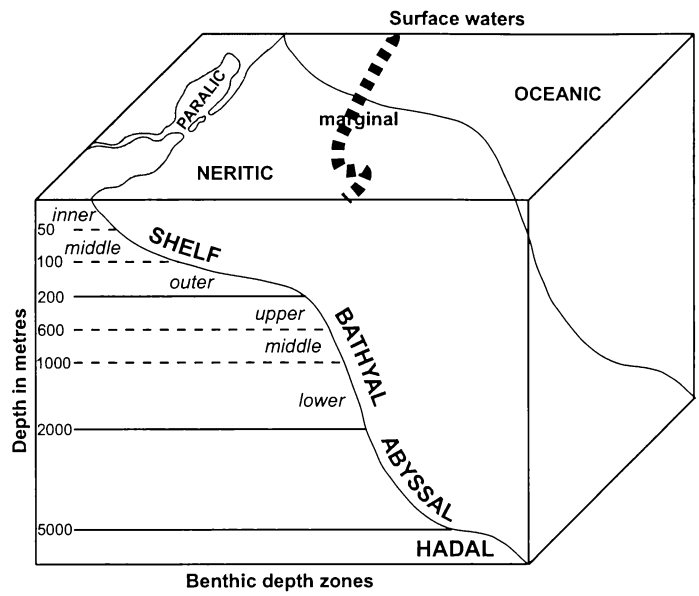
A total of 174 samples from the four wells were picked for full identification, and a further 204 for planktic-benthic ratios. Specimens have been identified to species level where possible, and to at least generic level. Those specimens which are highly degraded through (probably) transport and subsequent abrasion, and identified as *benthic* sp. *indet.* (species indeterminable) are not included in any analyses. While outcrop or dredge samples are not limited in size, cuttings samples by nature are limited in this way (refer Chapter 3, section 3). It was therefore not possible at times to obtain 100 specimens. Where total numbers for particular samples are very low (particularly benthic specimens), these samples have been excluded from analyses. Taxonomic classification follows that of Loeblich and Tappan (1987). Refer to Appendix 3c for systematics. All depths pertaining to well site samples are given in drillers depths (m bKB).

6.3 Paleoenvironmental assessment

6.3.1 Planktic foraminifera as an indicator of oceanicity

Planktic foraminiferal abundance is primarily used in this study as a general indicator of oceanicity (i.e., oceanic versus neritic watermasses). These watermass classifications reflect the ratio of planktic to benthic foraminifera, where the proportion of planktic taxa tends to increase seaward. Other measures frequently used to help determine paleo-oceanicity above a site are test size, degree of encrustation, and diversity. While high planktic numbers have been documented in shallow water (nearshore) situations (e.g., Hayward, 1985a), Murray (1976) demonstrated that away from open oceanic conditions, there is an observable reduction in the planktic-benthic ratio, a progressive decrease in the size of planktonic tests, and a reduction in diversity of the planktic assemblage. A more recent study, collecting samples and data from numerous sites west of Taranaki, indicate that planktic percentages show a distinct increase with water depth, with lowest percentages found close to shore, and highest percentages at the deepest stations (over 2000 m; B. Hayward and H. Grenfell, unpublished data, 2001).

In general, planktic foraminifera are most abundant in the euphotic zone (upper 200 m of oceanic water) where the greatest concentration of food occurs (particularly in areas of divergence and upwelling, and within major currents or boundary currents; Murray, 1991b; 1995). They are not tolerant of hypersaline, fresh or brackish water and do not thrive in the more turbid, neritic water of continental shelves and coastal zones (Boltovskoy and Wright, 1976; Hayward, 1986; Gibson, 1989). As oceanic circulation patterns are somewhat controlled by bathymetry (Murray, 1991a) it follows that nearshore sediments will tend to contain few planktic specimens, while deep oceanic sediments (above the CCD) often contain abundant planktic accumulations. Benthic populations, on the other hand, decrease rapidly away from coastal zones (Berger and Diester-Haass, 1988; Gibson, 1989). This means that planktic percentage can be used with some success to indicate whether a site was beneath open water, near land or enclosed, or whether these conditions changed over time, and even the direction to the shoreline, and possibly the nature and location of barriers to oceanic circulation.



***Benthic depth zones*:**

littoral = intertidal			
shelf		0-200 m	
	inner		0-50 m
	middle		50-100 m
	outer		100-200 m
bathyal			200-2000 m
	upper		200-600 m
	middle		600-1000 m
	lower		1000-2000 m
abyssal			2000-5000 m
hadal		> 5000 m	

Oceanicity:

oceanic - indicative of open water conditions

neritic - indicative of enclosed or conditions near land

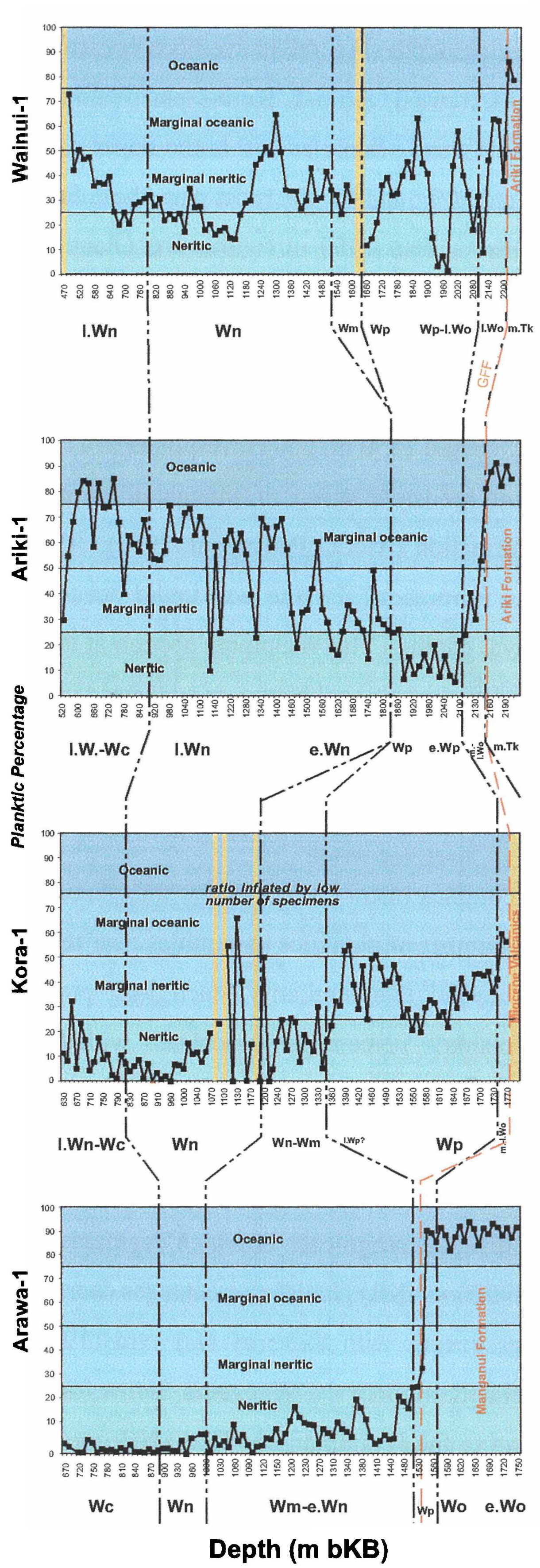
Fig. 6.1: Terminology adopted for benthic depth zones and surface watermasses. From Hayward et al. (1999).

As shown by Fig. 6.1, neritic water typically overlies much of the shelf, including enclosed bays cut off from oceanic circulation, while oceanic water occurs seaward of the shelf edge (Hayward, 1986). Between these two watermasses there occurs a broad 'intermediate' zone, through which the planktic ratio steadily increases. Hayward et al. (1999) have divided this zone into two distinct watermasses; a marginal neritic zone (25-50% planktics) and a marginal oceanic zone (50-75% planktics) while neritic waters contain less than 25%, and oceanic waters typically contain greater than 75%, planktics.

6.3.1 (i) planktic-benthic ratios

Simple plots of planktic percentage versus drilling depth (Fig. 6.2) have been constructed for each of the four well sections studied in detail (Arawa-1, Ariki-1, Kora-1 and Wainui-1). Of the four wells, Arawa-1 shows the most clear-cut trend. Samples within the upper part of the Manganui Formation display values greater than 80%, and for the most part, these lie around the 90% mark. In contrast, data points shallower than 1550 m (within the Giant Foresets Formation) are consistently less than 30%, dropping to less than 5% by 950 m bKB. These values are consistent with those obtained by Crundwell et al. (1992) for a sparser data set. The dramatic decrease in planktic percentage between the Manganui and Giant Foresets Formations occurs over a very narrow interval (between 1540 and 1550 m) and corresponds to a change in sediment type, from slightly sandy siltstone with sandstone stringers (Manganui Formation) to the massive siltstone of the Giant Foresets Formation (ARCO Petroleum, 1992). This trend indicates that during the Opoitian (Manganui Formation) the site was under oceanic water conditions and probably at mid to upper bathyal depths. The Waipipian section at this site is very thin (and may include a condensed interval or depositional hiatus, as suggested in Chapter 3) and it is during this interval that the marked reduction in planktic numbers occurred, although there is no significant change in paleodepth.

Ariki-1, Kora-1 and Wainui-1 display a more complex history of oceanicity than Arawa-1. All wells show an overall up-hole decrease in the number of planktics to benthics near the base of the Giant Foresets Formation. However, beyond that similarity, each well presents an individual history. Ariki-1 has very high planktic percentages associated with the Ariki Formation (greater than 80%), which decrease markedly (to <20%; Fig. 6.2) between the top of the Ariki Formation and the base of the Giant Foresets Formation. However, unlike Arawa-1, planktic percentages at this site then *increase* up-hole, with the increase punctuated at regular intervals by a number of obvious spikes and dips, the magnitude of which decreases up-hole. In the last 80 m of the sampled interval (600 to 520 m bKB) planktic percentages decrease from approximately 80%, to 30%. These findings concur with those of Hayward (1986), who suggested neritic to marginal neritic conditions throughout the Wanganui Series for this site, with perhaps more fully neritic conditions during the Waipipian and Early Nukumaruan, and more oceanic conditions during the Opoitian and Late Nukumaruan.



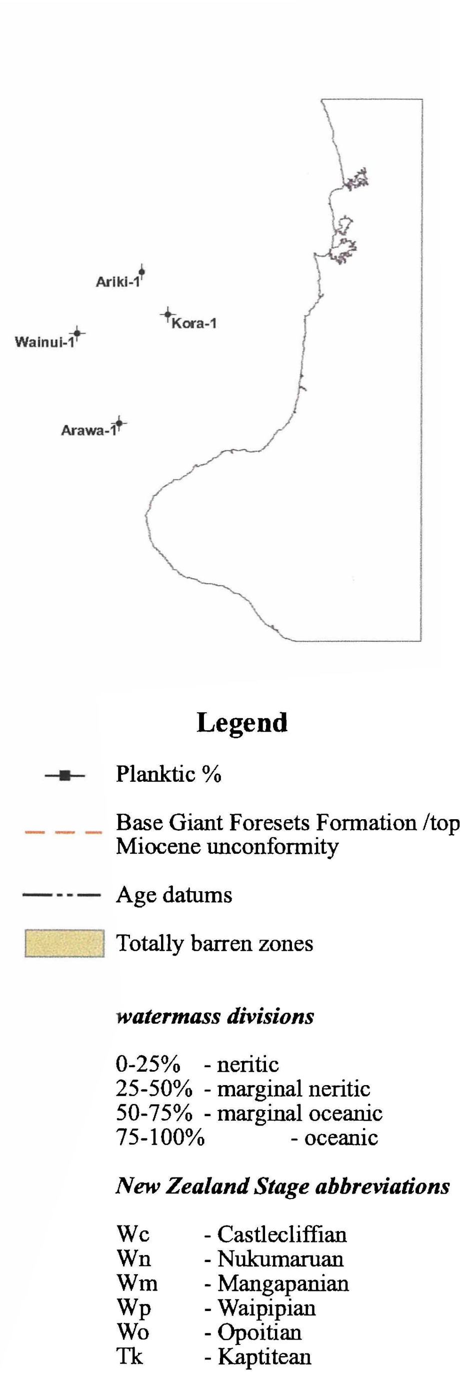


Fig. 6.2: Planktic percentage as an indicator of oceanicity. Water mass zones are after Hayward et al. (1999). Note the large decrease in planktic percentage observed between the Ariki Formation (or Manganui Formation) and the Giant Foresets Formation, at Wainui-1, Ariki-1, and Arawa-1. Planktic percentage fluctuates dramatically throughout the Plio-Pleistocene.

The low percentage of planktics observed during the Waipipian to Early Nukumaruan at Ariki-1 may reflect marked sea-level lowering, or the waning of oceanic/nutrient rich currents above the site. The lower planktic percentages may also be partly attributed to the initiation of terrigenous sedimentation during the Late Opoitian, after the site had been starved of terrigenous sediment through the Kapitean and Early/mid Opoitian, effectively 'diluting' the planktic signal with an increase in transported benthic species. The increasing planktic percentage throughout the Nukumaruan may merely reflect a slow change back to equilibrium with the overlying watermass. While it is not thought that the site was suddenly closed off from oceanic circulation, a change in regional current patterns might have affected this site, resulting in lower nutrient availability and therefore less food that could be utilised by planktic foraminifera. Hayward (1990, pg. 79) suggests that 'uplifting of land or shelf areas' in the Southern Inversion Zone of Taranaki Basin may, by the Late Miocene, have diverted northward-moving oceanic currents away from the coast in eastern Taranaki Basin.

A similar mechanism could be applied to the prograding slope, deflecting nutrient-rich currents away from the northern part of Taranaki Basin. These currents may have slowly returned by the Late Nukumaruan (with a concurrent increase in planktic percentage), only to be displaced again in the Late Nukumaruan by progradation of the shelf margin. This mechanism may also account for the very neritic conditions seen at Arawa-1; instead of very shallow conditions, planktic percentages may in fact be indicating that nutrient-rich currents were also being directed away from this site.

The Late- and post-Miocene section in Wainui-1 also displays a record of fluctuating planktic percentage, from >90% in the Ariki Formation, to a few percent in Giant Foresets sediments. However, unlike Ariki-1 and Arawa-1, the decrease in planktic percentage across the Ariki Formation/Giant Foresets Formation boundary is not nearly so dramatic. This may reflect the more basinward position of this site, having been at bathyal depths until the Late Nukumaruan (refer section 6.3.2 (ii)). Throughout most of the Plio-Pleistocene history at Wainui-1, planktic percentages are lower than expected, with the majority of samples falling within the neritic to marginal neritic zone, rather than the marginal oceanic to oceanic zone where the well was sited. Hayward (1984) also notes this anomalous signature, and suggested that given the present seafloor depth at this site, neritic conditions were not present at this site even during glacial low sea levels.

Kora-1 did not intersect the Ariki Formation. Instead, Late Opoitian sediments rest unconformably on undifferentiated Miocene volcanics. It is therefore not possible to determine whether the same trend of a sudden decrease in planktic percentage up-hole is present at the base of the Giant Foresets Formation. However, this site does display the same overall uphole shallowing watermass trend, with planktic percentage ranging between 50% at the local base of the Giant Foresets Formation, to just a few percent near the top of the sampled interval, with numerous fluctuations. Much of the data suggests the site was overlain by predominantly neritic to marginal neritic watermasses through the Pliocene and Pleistocene, with two brief intervals during which marginal oceanic watermasses were present (Late Opoitian and Late Waipipian). It is assumed from both the evidence gained from samples above the Miocene volcanics, and the inferred paleobathymetry of sediments below the volcanics, that the volcanic pile itself accumulated at bathyal depths (Hayward and Strong, 1993), probably under a marginal oceanic to oceanic watermass. Kora-4, located approximately 5 km to the southwest, mirrors the shallowing trend observed at Kora-1. Taxa and planktic percentages for the post-Miocene section in Kora-4 suggest an upper bathyal to inner/middle shelf environment through the upper Opoitian to lower Nukumaruan, and inner to outer shelf depositional environments for the remainder of the Nukumaruan and during the Castlecliffian (ARCO Petroleum NZ, 1985; King and Thrasher, 1996).

Other well sections in offshore Taranaki Basin show similar Late- and post-Miocene patterns in planktic-benthic ratios to those displayed for the well sections discussed above. Taimana-1, ~14 km to the southwest of Arawa-1, exhibits the same trend as observed in Arawa-1, with planktic percentages decreasing sharply from 75% during the Opoitian, to 25% during the Waipipian, and less than 10% during the Mangapanian to Castlecliffian (Diamond Shamrock Oil Co (NZ) 1984). While the decreasing planktic ratio does not correspond to the base of the Giant Foresets Formation (although a thin carbonate unit is recorded at the base of the formation), the timing of this event is consistent with the marked uphole decrease noted at Arawa-1. At Tane-1, situated 75 km to the south-southwest of Wainui-1, a decrease in planktic percentage up-section of the Late Kapitean-aged Ariki Formation is also observed, suggesting that this site also experienced shallowing watermasses through the Plio-Pleistocene (Shell BP Todd Oil Services, 1976).

Tangaroa-1 and Te Kumi-1 both exhibit planktic percentages and associated watermasses that vary uphole from strongly neritic to marginal neritic, with occasional peaks indicating marginal oceanic watermass conditions (Waghorn et al., 1996). At Tangaroa-1, planktic percentages for the Late Tongaporutuan to Early Kapitean indicate marginal oceanic to oceanic watermasses, and lower bathyal depths, while high planktic percentages (greater than 50%) are observed during the Opoitian. Fully oceanic watermass conditions prevailed at Te Kumi-1 until at least lower Waipipian. Unfortunately at this site, *in situ* fauna in upper Tongaporutuan to lower Kapitean sediments are heavily masked by down-hole cavings, and Waghorn et al. (1996) were unable to determine watermass conditions during this stage, although they postulate that depths were probably mid to lower bathyal.

The sedimentary sequence at both Awatea-1 and Mangaa-1 includes a very thick post Miocene sequence, encompassing the Mangaa Formation (a series of stacked basin floor fans) and the Giant Foresets Formation. Previous investigations indicate that both of these sites display similar trends, with watermass conditions shallowing from mid to lower bathyal through the Late Miocene and into the Opoitian, to mid/outer shelf depths by the Castlecliffian Stage (Strong et al., 1996; Waghorn et al., 1996). Neither site records a marked decrease in planktic percentage between Miocene and Pliocene strata. Watermass conditions for Mangaa-1 are summarised in Fig. 6.15 (section 6.2.3(iv)). A similar trend of neritic to marginal neritic/ marginal oceanic watermass conditions, and inferred shallowing trend from lower to mid bathyal (Late Tongaporutuan to Early Kapitean) is also evident at Kahawai-1 (Waghorn et al., 1996), although this site was undergoing mild uplift through the Pliocene and Pleistocene. Plio-Pleistocene erosion has subsequently removed latest Miocene to Late Pleistocene sediment from sites further eastward (e.g., Turi-1, Awakino-1, Mokau-1, and Pluto-1), which means that paleoenvironments can only be inferred.

In general, all wells in the study area, other than those very close to the modern-day shoreline, suggest that during the Late Miocene, the entire basin was under marginal oceanic to oceanic watermasses, with inferred mid to lower bathyal depths. Watermass conditions changed from oceanic to marginal oceanic during the Late Miocene to Opoitian, to marginal neritic and neritic in the Waipipian. However, while paleodepths shallowed, benthic faunas (6.3.2) suggest that this shallowing was not as marked as watermass conditions indicate.

6.3.1 (ii) Diversity

Planktic faunal associations tend to become more diverse with increasing water depth, and also reflect changing sea surface temperatures (warmer water generally equates to greater diversity). Planktic faunal associations can therefore also be used as an indicator of paleo-oceanicity. Simple diversity (number of species) is used most commonly for planktic foraminifera, although some authors (e.g., Ottens and Nederbragt, 1992) have used a combination of simple diversity, Shannon-Wiener information function, and a measure of equitability, to characterise modern ocean environments. As this study is also using benthic foraminiferal associations to interpret paleobathymetry, and due to the partial pressure-dissolution of (particularly) planktic specimens making it often difficult to identify some individuals beyond the generic level, only simple diversity is discussed in this section.

Arawa-1 and Kora-1 (Fig. 6.3) display an overall upward (uphole) decrease in diversity, concomitant with the upward shallowing trend indicated by the decreasing planktic ratio. It is unclear why there should be a diversity peak at 960 m in Arawa-1 – as for many samples in the Giant Foresets Formation, planktic numbers are extremely low, yet this sample contains a greater number of species. As this peak involves only one sample, not much weight can be placed on it. Diversity at Kora-1 is much higher below the barren/nearly barren zone than above this interval.

The shallower water sites Arawa-1 and Kora-1 display a trend of decreasing planktic diversity with decreasing water depth from the Late Pliocene onwards. This is in keeping with published diversity trends and shallowing attributed to progradation of the continental margin past these sites. However, the deeper water sites Wainui-1 and Ariki-1 display the opposite trend, with increasing planktic diversity with shallowing water depth. Therefore an alternative explanation for these deeper sites must be invoked. The diversity trend could be related to change in nutrient supply, which in turn relates to paleocirculation. General circulation patterns in the southern Tasman Sea are inferred to have remained relatively stable since at least the Miocene (Nelson and Cooke, 2001; Cooke, 2002). However, the coastal circulation patterns must have been influenced by the changing sedimentary regime and rapid progradation of the continental margin (Chapter 5).

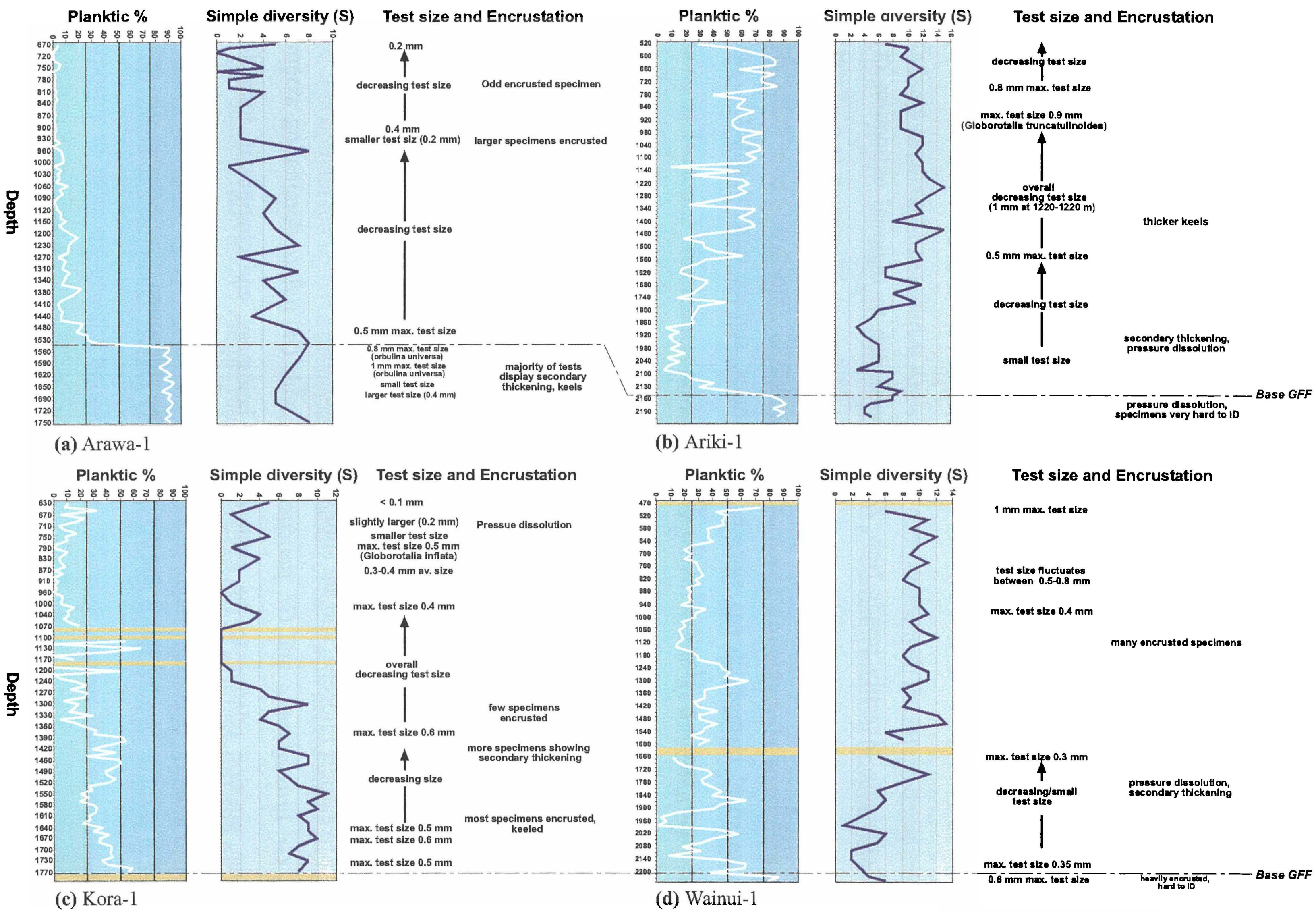


Fig. 6.3: Summary figure displaying planktic diveristy (*S*), test size with depth, and secondary calcite thickening. Light brown colour (▭) indicates barren zones

During the Late Miocene, the continental margin was south of the study area, sediment starvation was prevalent across much of the northern part of the basin, and all of the well sites were bathed by bathyal water depths. As the continental margin prograded northward, it is proposed that the margin deflected the Tasman Current seaward, thus reducing the nutrient supply to the deepest sites and consequently reducing planktic productivity, which accounts for the initially low species diversity observed in Wainui-1 and Ariki-1. Subsequently, these deeper sites adjusted to changing nutrient levels associated with suspended sediment influx (= nutrients), which accounts for the increasing species diversity and planktic percentage in these wells during the Late Pliocene to Pleistocene. In addition, a number of other mechanisms could also be brought into play to explain the observed diversity trends. These include:

1. changing diversity trends New Zealand-wide due to regional extinction of many globorotalid species during the Late Miocene through Waipipian (e.g., Scott et al., 1990);
2. seasonal changes in sea surface temperatures, surface current patterns, and food supply (see e.g., Hayward et al., 2002).

Seasonal variations, such as discussed in Hayward et al. (2002) are impossible to quantify in this study because of inherent resolution difficulties when dealing with cuttings samples from boreholes. Extinction of numerous planktic species during the Late Miocene to Opoitian (Fig. 3.2) may account for some of the observations. For example, at both Ariki-1 and Wainui-1, species such as *Gr. miotumida*, *Gr. mons*, *Gr. pliozea*, *Gr. puncticulata* and *Gr. p. sphericomiozea*, prevalent in Ariki Formation samples, become extinct during the latest Miocene to earliest Pliocene, and are abruptly replaced in the Giant Foresets Formation by *Gr. inflata*, *Gr. crassaformis*, and *Gr. puncticuloides*, with *Gr. inflata* dominating assemblages. However, position relative to the contemporary landmass, and changes in local (coastal) currents during cyclical(?) changes in sea level, are probably the most likely factors governing changes in planktic-benthic ratios and diversity in the study area.

6.3.1 (iii) Test size and encrustation

Because optimum growth conditions of planktic foraminifera occur in truly oceanic water, maximum test size increases with increasing distance from land (Murray, 1976; Hayward, 1986; Lohmann, 1995). Neritic-dwelling specimens, on the other hand, are usually small, and may be composed entirely of juveniles (Murray, 1976). Hence the maximum size (taken as the largest specimen in a sample irrespective of genus or species) can be useful in indicating relative water depth. The same applies for the degree of encrustation of some individuals. Encrustation (thickening of tests by addition of secondary calcite) tends to occur primarily at bathyal depths, with thinner shelled specimens more apparent in shallower waters (Hayward, 1986). The appearance and dominance of keeled globorotalids (e.g., *Gr. truncatulinoides*, *Gr. miotumida*) are also suggestive of deeper water, as these live at depths >200m, and will therefore be deposited at bathyal or greater depths. A summary of test size and degree of encrustation is included on Fig. 6.3.

The relatively small test size noted at Arawa-1 is in accordance with the inferred watermass and paleodepth. Overall test size decreases uphole, with a minimum of about 0.2 mm between 670 and 810 m, although several lower sample intervals (e.g., 1640 m) are also comprised of relatively small tests (< 0.4 mm). *Orbulina universa* appears to obtain the maximum test size of any species, with a ~0.8 mm specimen noted at 1570 m, and specimens approaching 1 mm in diameter at 1620 m. This reflects a more oceanic environment as indicated by watermass conditions. However, pressure dissolution, presumably associated with the thickness of overlying strata, becomes more pronounced with depth, and appears to affect thinner-walled *O. universa* more conspicuously Secondary calcite encrustation, thickened keels, and evidence of pressure dissolution are prevalent until approximately 1540 m bKB, although the rare encrusted specimens are noted above this interval (e.g., 750 m). Test size reduces between 1540 m and 1460 m, increasing in size again between 1430 and 1180 m, and between 920-850 m (individual specimens up to ~0.4 mm). *Globorotalia inflata*, prevalent in many samples, is observed to display a concomitant reduction in the degree of inflation of tests with decreasing test diameter.

Taxa in the deepest (downhole) sample intervals at Ariki-1 (>2150 m) are difficult to identify beyond generic level because of the prevalence of calcite encrustation and pressure dissolution effects (distortion of specimens). Heavy iron staining of specimens noted at 2060 m suggests some exposure to more oxygenated bottom waters. Varying degrees (more prominent with depth downhole) of secondary calcite thickening, pressure dissolution and thickened keels are a feature of many planktics until about 1400 m bKB. Test size generally decreases uphole, particularly *Gr. inflata*, although large specimens of this taxon are recorded at 1200 m (~1.0 mm), and 1140 m (~0.5 mm), and *Globorotalia truncatulinoides* reaches a maximum diameter of 0.9 mm at 940 m bKB. As with Arawa-1, many fluctuations in mean test diameter occur, particularly between 2100-1560 m, 1520-860 m, 780 m, and 700-520 m.

As with other sites, deeper sample intervals at Kora-1 (below 1400 m bKB) are affected to some degree by secondary thickening coupled with pressure-dissolution. Many specimens display signs of deformation below 1650 m. More keeled specimens are also observed below 1400 m, with few specimens affected by pressure dissolution and secondary thickening higher in the sequence. Text size generally decreases uphole, with largest mean diameters recorded at 1680 m (up to 0.6 mm), 1540 m (0.5 mm), 1380 m (up to 0.6 mm; *Gr. inflata*), and 800 m (0.5 mm; *Gr. inflata*). Test size decreases between 1510-1530 m (<0.4 mm), increases to 1380 m (0.4-0.5 mm), and decreases again to 1240 m (max. 0.3 mm). Average test diameters vary between 0.3 and 0.4 mm between 1030 m and 810 m, but are no greater than ~0.1-0.2 mm above 800 m.

On average, Wainui-1 records larger test sizes than at any other site, with a maximum text diameter of ~1.0 mm recorded at 480 m (*Gr. truncatulinoides*). Test size on average is larger in the deeper sampled intervals (~0.6 mm at 2220 m), complemented by a prevalence of faunas displaying keels, secondary thickening and pressure dissolution. Below 1920 m, individuals become harder to confidently identify past generic level, and most specimens below 2200 m are heavily encrusted and display some degree of partial dissolution. Occasional samples higher in the sequence also show evidence of secondary thickening, particularly at 1120 m, where approximately half of the *Gr. inflata* are heavily encrusted. While the odd sample lower in the sequence is characterised by small test size (e.g., 2210-2180 m (0-.35 mm), and 1840 m (0.3 mm)), overall test diameter decreases up-hole. Test size decreases then increases slightly after the barren interval between 1620-1640 m, and tends to vary markedly with decreasing sample depth. For example, large individuals are observed at 940 and 820 m (0.5 mm), 540 m (0.6 mm), and 560 and 500 m (0.7 mm). *Orbulina universa* are the largest planktic foraminifera recorded between 900-560 m, varying between 0.5 and 0.8 mm, and *Gr. inflata* reach a maximum size of 0.5 mm at 940 and 820 m.

Overall, simple diversity, test size, and degree of secondary calcite thickening, display good correlation with planktic-benthic ratios. Largest test size corresponds to more oceanic watermasses, while overall test size decreases up-hole, as does the degree of encrustation displayed by total planktic assemblages. The general pattern is one of decreasing oceanicity with (inferred) decreasing depth, punctuated by periods of increasing oceanicity.

6.3.2 Benthic foraminiferal assemblages - paleoenvironmental indicators

Paleoenvironmental interpretations of shelf to basin marine environments, using foraminifera, are largely based on the taxonomic composition (i.e., an assemblage or biofacies) of benthic foraminiferal faunas, with key indicator (depth-restricted) taxa used to identify successively deeper or shallower environments. Hayward (1986a) discussed the need for an 'all in one' guide outlining the criteria useful in determining paleoenvironments from foraminiferal data. That study details a number of techniques for qualitatively identifying depositional environments, collated from the study of fossil occurrences in several Taranaki Basin drillholes, and integrated with modern records of the depth distribution of many of the taxa used, or closely related taxa. Because many extant species occur in the fossil record, it is possible to use the depth ranges of the modern-day relatives and apply these to paleoenvironmental/paleo-geographical studies. Hayward et al. (1999) builds upon the earlier (1986a) study, using more quantitative methods (e.g., correspondence analysis) for determining specific depositional environments of assemblages of faunas, rather than individuals. It provides a comprehensive account of the distribution of benthic faunas in New Zealand's modern shallow-water (shelf and paralic/brackish) depositional environments.

Qualitative interpretative methods are based on the traditional taxonomic principal of uniformitarianism (the present is the key to the past). This approach is based on the supposition that the environments occupied by fossil species are the same or similar to their taxonomically nearest living relatives. However, this key principal does not allow for ecological shifts. Because it is unlikely that an entire assemblage of benthic foraminifera would have changed their ecological niche simultaneously (Bosence and Allison, 1995) it is considered that the use of more quantitative statistical methods, and review of entire assemblages, rather than single species, will provide more precise paleoenvironmental interpretations. As such, interpretations of benthic faunas in this chapter utilise both the qualitative methods outlined in Hayward (1986a), and integrates these with the more quantitative methods discussed in Hayward et al. (1999).

Methodologies used for the interpretation of benthic fauna here include depth calibration of specific bathyal-restricted taxa using seismic reflection profiles (6.2.3(ii)), taxonomic composition by suborders (6.2.3(iii)), and cluster analysis of both species and samples (6.2.3 (iv)). Species diversity (6.2.3(v)), and species dominance versus textural trends (6.2.3(vi)) are also reviewed.

6.2.3 (i) A brief introduction to the benthic foraminifera.

Benthic foraminifera are present in a wide range of marine environments, from brackish estuaries and inlets, to the deepest parts of the ocean. Depth itself is not perceived to be a limiting factor. However, a number of parameters such as temperature, light intensity, salinity, and density, all considered to impact on benthic distribution, are closely related to depth (Murray, 1992). The most important governing factors, although, appear to be oxygen and organic flux (e.g., Berger and Diester-Haass, 1988; Gooday, 1988; 1994; Van der Zwaan et al., 1999), with competition for food (Murray, 1992) a third parameter. Light intensity, temperature, substrate, and energy availability are responsible for major benthic foraminiferal assemblage changes from the photic zone towards deeper water (Murray, 1992; Van der Zwaan et al., 1999), whereas faunas on the slope and basins generally show distribution patterns related more to bottom watermass conditions and food availability. Areas beneath zones of upwelling or high surface productivity, in particular, result in high organic contents in the sediments below (Murray, 1992; Hayward et al., 2002), with resultant high concentrations of benthic foraminifera. On a global scale, a regular pattern of species distribution occurs, with some species typically occupying deeper water (adapted to low organic flux), while others proliferate at the shelf margin (require a high organic load to flourish; Van der Zwaan et al., 1999). On a smaller scale, the depth ranges of these species may vary, depending on available substrate type, food, and local bottom watermass properties (Murray, 1992).

Organic flux and oxygen availability

Murray (2001 pg. 3), states that it 'would be too simplistic to attempt to define all distributions in terms only of oxygen and food supply ... success of foraminifera must be attributed to their ability to tolerate a broad range of environmental conditions'. However, food is deemed to be a structuring parameter (Van der Zwaan et al., 1999), affecting species abundance, until oxygen limitation prevents the occurrence of a species. Most benthic foraminifera prefer the sediment-water interface as a place to live, since the most abundant resources are available there (bacteria or organic matter). Van der Zwaan et al. (1999) suggest that often more than 75% of benthic foraminifera live in the top cm of the sediment column (epifaunal and shallow infaunal; Corliss, 1991). Studies have shown that total benthic numbers at first increase proportionally in conjunction with increasing organic matter (nutrients). At a certain point, because oxygen is consumed in the natural transformation of carbonaceous compounds (Sen Gupta and Machain-Castillo, 1993), oxygen deficiency starts to become a limiting factor, leading to diminishing numbers (Van der Zwaan et al., 1999). Epifaunal benthic foraminifera tend to be the first affected, whereas the deeper infaunal groups, living under diminished oxygen conditions anyway, are more tolerant, and remain the longest (Van der Zwaan et al., 1999). When this oxygen-limiting effect occurs, the environment is said to be 'stressed'.

Opportunistic species

Opportunistic species are ones that can proliferate in the face of adverse environmental (stressed) conditions, such as low oxygen environments. Under the usual combination of oxygen depletion and organic matter enrichment of the substrate, the populations of some of these species (e.g., *Uvigerina*, *Evolvocassidulina*, *Cassidulina*, and *Bulimina*; Sen Gupta and Machain-Castillo, 1993; Gooday, 1994) expand significantly. In the opportunistic model, with increasing organic flux comes increased demand for oxygen. At some point this increased need for oxygen leads to lower oxygen contents at the surface-water interface. This forces species to tolerate low oxygen conditions if they are to exploit the rich food source. Those species that are able to profit from this situation do so by having the ability to reproduce rapidly, so that they reach extremely high numbers in a very short time. However, when the food supply decreases, it is the opportunistic species that are most affected (Gooday, 1994; Van der Zwaan et al., 1999). It is important to note, however, that many species that are said, or known, to be opportunistic, can be equally as frequent in eutrophic environments where the oxygen content is high.

Stressed environments can occur through several different mechanisms. Probably the most studied is that which occurs in areas of coastal upwelling or other areas of high productivity. In these areas, large amounts of organic carbon accumulate in bottom sediments, leading to oxygen depletion in bottom waters, and pore waters (Sen Gupta and Machain-Castillo, 1993). Schnitker (1994) considers that benthic foraminifera are unequivocal indicators of surface water productivity, where productivity is high. Mobility of substrate is another important mechanism. Marine environments that are subject to frequent environmental disturbances, such as shelf margins and slopes, may be rapidly colonised by opportunistic species (Murray, 1992; Alve and Murray, 1997; Alve, 1999). Environmental disturbance on the slope and basin floor may be a particularly important mechanism in creating stressed environments within the study area.

Benthic assemblages and their relation to sediment

Large quantities of sediment being reworked and transported from one environmental zone to another have two major effects on the fossil benthic assemblages later analysed:

1. areas of rapid sedimentation tend to smother existing communities to some degree, but also may bring high concentrations of nutrients (Alve and Murray, 1997). This provides an opportunity for species to utilise the rich food source and proliferate rapidly. These variable, stressed conditions may later manifest themselves in the stratigraphic record as intervals of simple diversity (i.e. low number of species) and high abundance (Murray, 1992);
2. reworking and transport from shallower regions to deeper regions results in mixing of shallow water faunas with deeper ones. This provides evidence of a paleoslope and shallower environments nearby. Absence of shallow water tests may conversely indicate a site is some distance from shallower areas, or that the area was isolated from sediment input (e.g., bathymetric high; Hayward, 1986a).

Mechanisms for penecontemporaneous transport of shallow water faunas, and which may result in catastrophic disturbance of the environment, include sudden depositional events such as turbidity currents (e.g., Berggren, 1978), and subaqueous mass flows, such as debris flows and slumping (Hayward, 1986a; Alve, 1999), or volcanic deposits (Alve, 1999). In the 1979 study of Hayward and Buzas, samples that had mixed shallow and deep benthic faunas were considered to result from downslope sediment gravity flows. Apart from shelf margins and slopes, deep-sea conditions may also be unstable, with postmortem transport of tests taking place through small or large scale slumping, as well as sliding through the action of bottom currents (Hayward, 1986a; Murray, 1992). Disturbance to the sea floor can have varying degrees of destructive impact on benthic communities, including defaunation of the existing sediments, or establishment of new, unoccupied habitats (Alve, 1999). Azoic zones (i.e., zones that are barren of living faunas) may be created by the stripping of surface sediments during mass flow events, or may be due to an environmental change such as retreating ice fronts and oxygenation following anoxic events. Alve (1999) considers that marine transgression probably operates too slowly to provide a lag between exposure of new sea floor and colonisation.

6.3.2 (ii) Depth determination of key species

Benthic faunas at inner and mid shelf depths are generally more distinctive and stratified than those at greater (outer shelf and bathyal) depths. Resultant paleodepth assessments of shallower water faunas are therefore fairly accurate (±20-50 m). However, difficulties arise when trying to determine paleodepths within the 200-3500 m bathyal zone, as many taxa exist throughout this range. Accuracy limits in the outer shelf to uppermost bathyal zone vary by ± 50-200 m, while large accuracy limits of ± 200-1000 m are expected in the bathyal zone (Hayward, 1986). A number of studies have been undertaken in order to determine the present-day distribution of deep-sea benthic foraminifera, including Lewis (1979) and Hayward et al. (2001; 2002). While these studies have presented depth ranges of certain (common) benthic faunas in Hawke Bay and in the vicinity of the Chatham Rise, respectively, these ranges can change dependent on location. It is therefore important to clarify the depth ranges utilised in this study, particularly upper depth limits of certain key depth-restricted faunas.

Seismic reflection calibration – an independent method of determining paleodepth

Seismic lines provide a means of independently calibrating the uppermost stratigraphic ranges of specific diagnostic and bathyal-restricted benthic foraminifera at any one site, establishing an ordering of the upper paleodepth limits of the taxa. This method is outlined in Hayward (1986) and has been used on a number of wells in the Taranaki Basin (e.g., see Hayward, 1990; Crundwell et al., 1994) as well as in Hawke Bay (Scott and Crundwell, 1994).

The methodology behind utilisation of seismic lines as a calibrating tool relies upon identification of paleoshelf breaks. The shelf break at the time of deposition is defined as the change in slope between the topset and foreset parts of a single seismic reflector. Given that water depth to the shelf break is assumed to approximate the present day shelf margin (about 200 m), the depth anywhere along a reflector associated with a particular paleoshelf break can be estimated by calculating the depth above or below the shelf break for that location. Crundwell et al. (1994) use an interval velocity calculation of 2.44 km/sec derived from Taimana-1 velocity data to calibrate the depth of the shelf below the upper sequence boundary at that site, and have determined and calibrated the upper depth limits of significant species for a number of other wells including Ariki-1 Wainui-1 and Mangaa-1. This depth information has been summarised in Hayward et al. (1999) and is presented in Table 6.1.

Using the methodology outlined above, and the more detailed sampling intervals examined, this study has been able to determine the upper paleobathymetric limits of numerous key taxa for the well sections analysed (see Table 6.1), expanding on the number of taxa previously listed. This provides more accurate constraints on depth assessments of faunal associations (section 6.3.2) in northern Taranaki Basin. A summary of the species abundance of common and key depth-restricted benthic foraminifera with depth downhole, and interpreted paleodepths based on seismic calibration, is given in Appendix 3d (part b).

Table 6.1: Approximate calibrated upper water depth limit of selected key Neogene benthic foraminifera in offshore North Taranaki, and upper limits from comparative studies.

Benthic species	Hayward (2001, 2002), Chatham Rise	Hayward et al. (1999), Taranaki Basin	Crundwell et al. (1994), Taranaki Basin	Lewis (1979), Hawke Bay	This study		
					Arawa-1	*Ariki-1*	*Wainui-1*
Haeuslerella parri			200 ± 50 m		200 ± 50 m		>250 m
Hoeglundina elegans			250 ± 50 m	100 m	200 ± 50 m	>200 m	>250 m
Cibicides neoperforatus		300 ± 100 m			250 m	>250 m	>250 m
Bolivinita quadrilatera	250 m			50 m			
Karreriella cylindrica		300 ± 100 m			200 ± 50 m	500 ± 100 m	c.300 m
Pullenia bulloides			450 ± 50 m	1250 m	250 m	800 m	c.700 m
Sigmoilopsis schlumbergeri	750 m	600 ± 150 m		1250 m		550 ± 100 m	1000 ± 150 m
Stilostomella spp.						700 ± 100 m	400 ± 50 m
Pleurostomella spp.		700 ± 200 m				>1000 m	550 m
Pyrgo murrhina	700 m			350 m			
Eggerella braydi	750 m			600 m	300 ± 50 m	500 m	400 ± 50 m
Sipouvigerina notohispda		1000 ± 300 m				>1000 m	>1000 m
Gyroidina neosoldani		1200 ± 300 m				>1000 m	
Bulimina marginata f. *aculeata*	400			200 m	300 m	300 m	c.300 ± 100 m
Globocassidulina subglobosa					200 m	450 ± 100 m	

NB: Hayward et al. (1999) adapted from Hayward (1986a), King et al. (1993), and Crundwell et al. (1994). Taimana-1, Ariki-1, Wainui-1, and Mangaa-1 were analysed by Crundwell et al. (1994).

Upper depth limits for many species are comparable to those obtained by Crundwell et al. (1994) and presented in Hayward et al. (1999) for Taranaki Basin wells. These depth limits are often shallower than those determined for the same or similar species in Chatham Rise samples, highlighting the statement made in Hayward et al. (2002) that species ranges appear to be location-dependent. Upper depth limits appear to be slightly shallower for certain species at Arawa-1 (e.g., *Karreriella cylindrica*, *Pullenia bulloides*, *Eggerella braydi*) than at either Ariki-1 or Wainui-1, but for most species depth ranges are very similar.

Many of the benthic species listed in Table 6.1 have been noted to occur in approximately the same stratigraphic order downhole, but are diachronous (G. Scott, in Hayward, 1990; see also Appendix 3d, part b), which is related to the progradation of the foreset front. In other words, the sequence of highest occurrences is older in the south and east, and youngest in the north and west.

6.3.2 (iii) Taxonomic composition by suborders

Foraminifera are members of the Order Foraminiferida (Boltovskoy and Wright, 1976). The Order Foraminiferida is further divided into twelve suborders (after Loeblich and Tappan, 1987). These are: Textulariina, Rotaliina, Miliolina, Lagenina, Allogromiina, Fusulinina, Globigerinina (includes the planktic species used in this study), Involutinina, Silicoloculinina, Spirillinina, Carterinina, and Robertinina. Of the twelve, only the first four (most dominant in the fossil and modern record) are used in this study. Ternary plots of Rotaliina plus Lagenina (hyaline tests), Miliolina (porcellaneous tests), and Textulariina (agglutinated and siliceous tests) have a limited, but rapid, use in assessing possible paleoenvironments (Hayward et al., 1999). Faunas that contain more than 20% Miliolina indicate normal marine, inner shelf environments, whereas faunas dominated by Textulariina (>50%) tend to be either from brackish water, or from below the Carbonate Compensation Depth (CCD). Rotaliina dominate most faunas, but this may indicate environments anywhere from slightly brackish, to upper abyssal depths. Lagenina, if taken separately, can indicate outer shelf or upper bathyal environments where this subgroup occurs in numbers greater than 30% of the fauna. Generally, this suborder occurs in low numbers, and an analysis of all four wells shows that rarely do Lagenina percentages constitute more than 5% of the total fauna. Percentages of each suborder are listed in Appendix 3d, part a.

Hayward et al. (1999) suggest that using suborders to indicate paleoenvironments is of limited value because of the inability to discriminate between the majority of faunas that live in normal marine conditions at shelf and bathyal depths. As such, this technique was applied to the four wells as a gross comparative method only. Relative percentages of each of the suborders for each well were plotted against depth (Fig. 6.4) and are compared against a ternary diagram of environmental distribution of the four suborders (Fig. 6.5; after Hayward et al., 1999). Figure 6.6 shows the relative abundance and distribution of Textulariina, Miliolina, and Lagenina, as well as common genus and species depth distributions. Rotaliina is not included on this figure as the depth range for this suborder is so broad. Depth range of Rotaliina with faunal contents of 40-100% (of assemblage) is 0-50 m, and c.70-100% is 50-c.5000 m (Hayward, 2001, pers. comm.).

As Fig. 6.4 illustrates, all wells show a predominance of Rotaliina (plus Lagenina). At Arawa-1, Kora-1, and Wainui-1, values only occasionally dip below 75%, although decline with decreasing depth up-hole. At Ariki-1, above 860 m, Rotaliina values are consistently below 75%, with values ranging between 60 and 12%, and dominance alternates with the suborder Miliolina. In no well does Textulariina dominate, though in all wells, Textulariina percentages decrease with decreasing depth uphole, tending to be highest below the Giant Foresets Formation.

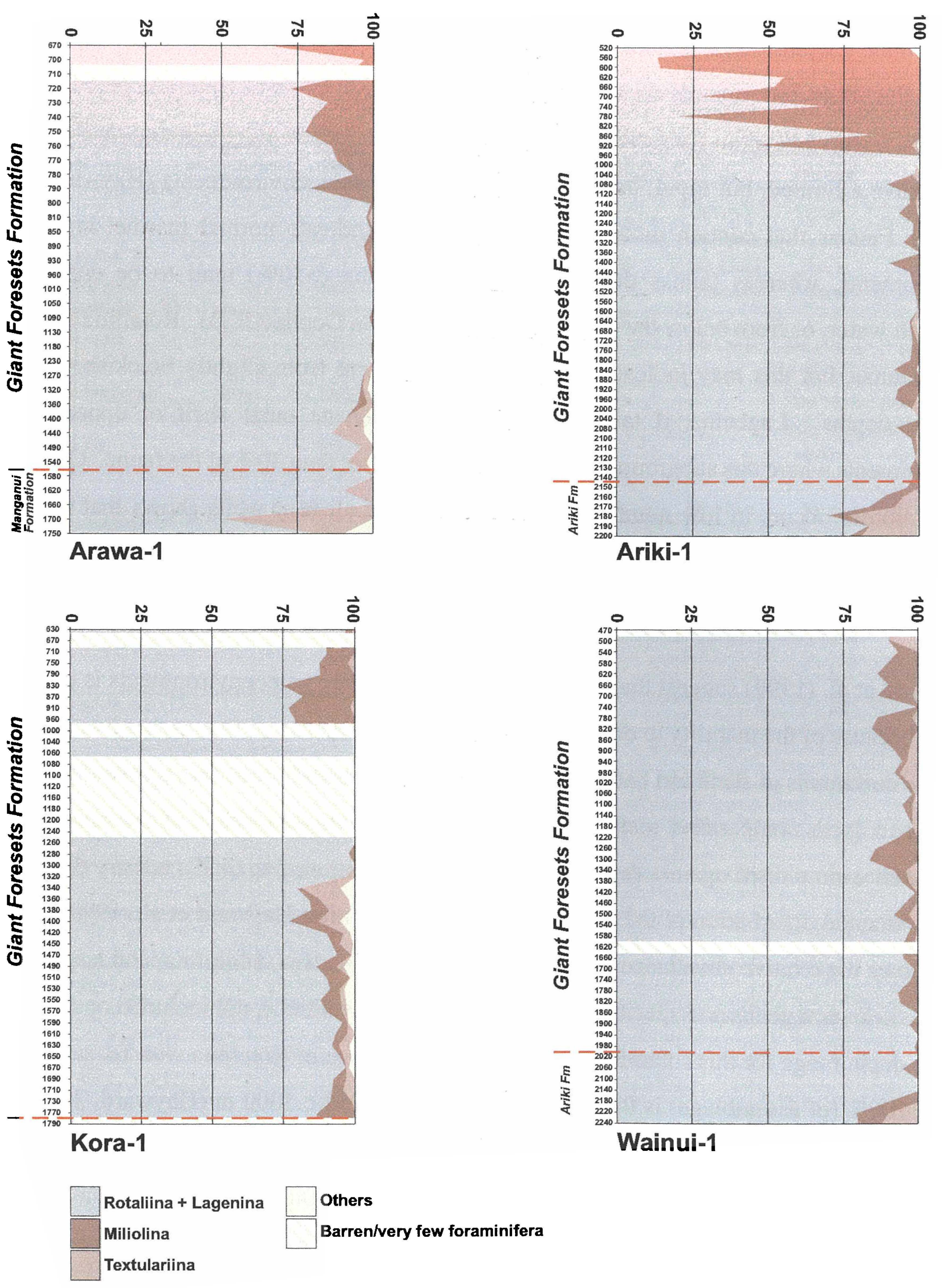


Fig. 6.4: Cumulative percentage plots of foraminiferal suborders for individual well sections. Note the dominance of Rotaliina (plus Lagenina).

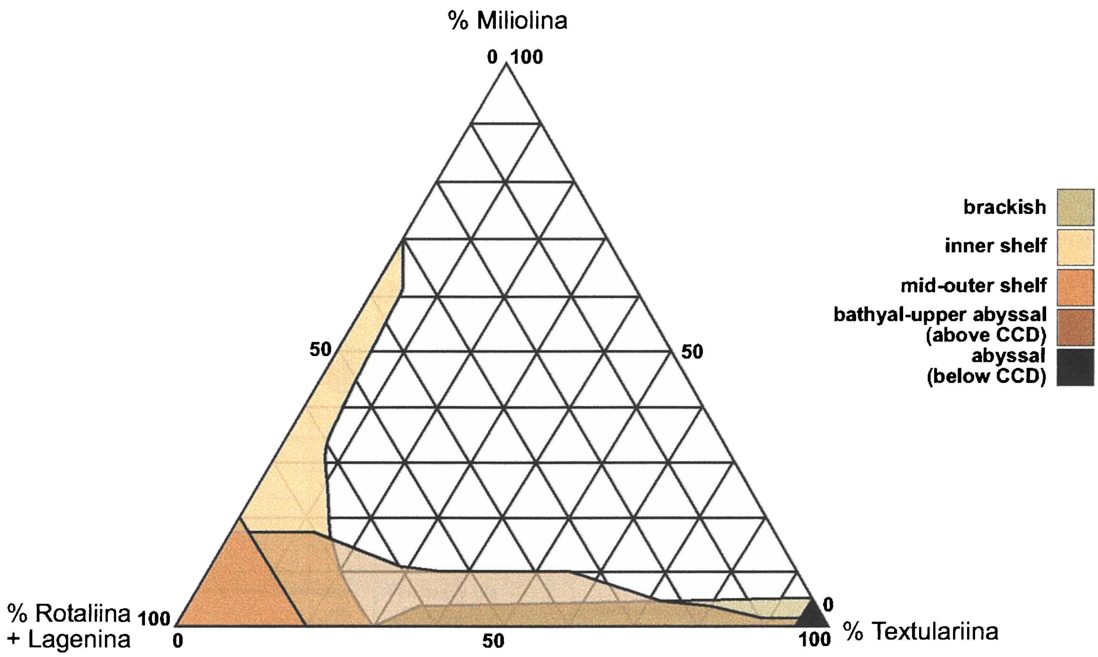


Fig. 6.5: Ternary diagram (after Hayward et al., 1999) demonstrating the environmental distribution of the four major suborders (Miliolina, Rotaliina, Lagenina, and Textulariina).

While species belonging to the Lagenina suborder are uncommon in these wells, and as such are lumped together with Rotaliina, they do reflect the relative position of each well site through time, with low but persistent numbers occurring throughout the Giant Foresets Formation at the more basinward Wainui-1 site, and values consistently greater than 5% (and very occasionally up to or just over 20%) below the Giant Foresets Formation in both Arawa-1 and Ariki-1 (refer to Appendix 3d). Although the dominantly high Rotaliina values can suggest environments anywhere from brackish to upper abyssal depths, the uphole increasing trend in Miliolina percentage, and decreasing Textulariina percentage for all wells, indicates a general shallowing upward trend, consistent with planktic foraminiferal trends. Low Textulariina percentages suggest that at no time over the sampled interval were any of these sites at either neritic/paralic depths, or at depths greater than the CCD.

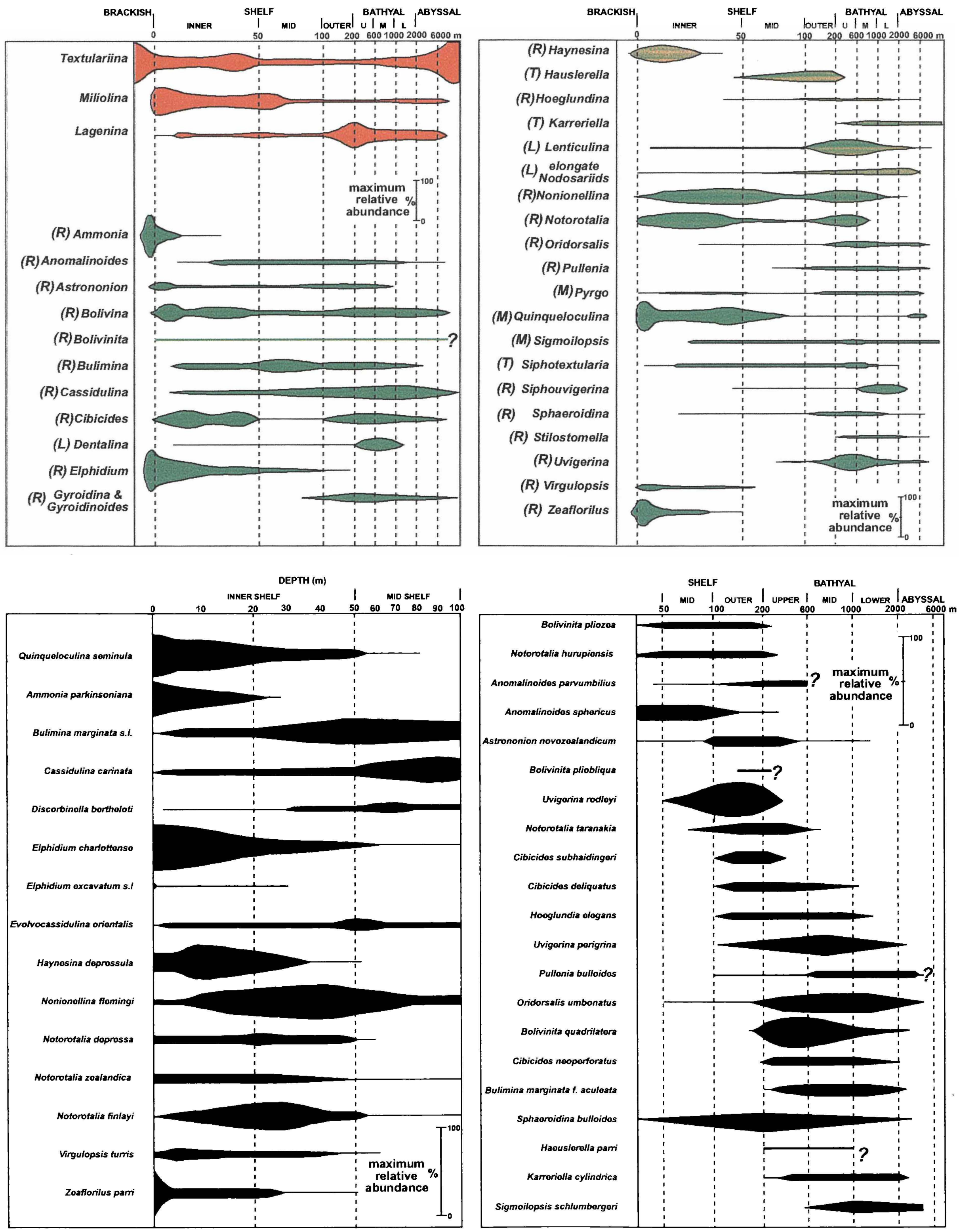


Fig. 6.6: Range and relative abundance charts of major suborders (red), genus (green) and species (black). Bracketed letters next to each genus relate to the suborder each genus belongs (L = Lagenina; M = Miliolina; R = Rotaliina; T = Textulariina). Shallower water species from Hayward et al. (1999). Deeper water species from Lewis (1979) and Hayward (1986a).

6.3.2 (iv) Taxonomic composition by genus and species: faunal assemblages

Taxonomic composition of the benthic fauna is the most precise method available for paleoenvironmental interpretation (Hayward, 1986a). Hence analyses of benthic faunal associations is a key technique used in this study to determine depth ranges and paleoenvironmental niches inhabited by specific taxa. This is done through the integration of the quantitative techniques of cluster analysis dendrograms, and the use of present day ecological ranges of extant species and inferred range of extinct taxa. Hayward (1986a) considers that the ecological ranges of present-day families, genera or species of New Zealand foraminifera are reasonably well known, and that only a few genera or species have drastically changed their environmental preference through time (mainly before the Neogene). Additional studies over the following decade (e.g., Hayward et al., 1997a,b; 1999) have refined knowledge about the distribution and habitats of shallower water taxa, while more recent work (e.g., Hayward et al., 2001; 2002) has been directed more towards deeper water taxa. Upper limits of depth-restricted fauna are primarily taken from Lewis (1979), Hayward (1986a), Van Morkhoven et al. (1986), and Hayward et al. (1999; 2001; 2002), as well as the seismically-calibrated depths presented in Table 6.1.

Cluster analysis dendrograms

Cluster analysis is a statistical multivariate analytical method that involves objects of a study (in this case, benthic species or genera) being divided into distinct groups, based on the characteristics of the objects (here, depth down-hole). Each object starts as a separate entity, and is then successively combined with the most similar objects until all are in a single, hierarchical group. This results in a dendrogram that displays the most similar cases linked most closely together (Kovach, 1999). This grouping enables the delineation of depth-related species assemblages, which can then be used singularly or in combination to identify the environment of deposition and/or indicate addition of reworked or transported faunas, at any sample depth (down-hole).

Cluster analysis was performed for each of the four control wells; Q-mode (sample by sample) analysis using Bray Curtis similarity coefficient was done on samples with more than 40 specimens. Because Q-mode cluster analysis provides no information on the composition of the species assemblages (Denne and Sen Gupta, 1991), R-mode (species versus species) analysis using modified Morista coefficient of similarity, was run on species with more than 10 individuals (over the entire sampled sequence). Cluster analysis was carried out using the computer program Multi Variate Statistical Package, version 3.1 (MSVP v. 3.1; Kovach, 1999). See Kovach (1999), Beals (1984), and Ter Braak (1986) for further discussion on the methodology behind this multivariate analysis technique. As the size of splits, and the number of benthic faunas picked, vary from sample to sample and well to well, counts were standardised by converting to proportions of the total benthics per sample. Species and faunal sample associations have been selected after scrutiny of results, and depth assessments have been based on the total faunal content, plus key environment-specific indicator species. This results in a more robust analysis, diminishing the effects of more rare species, while placing more weight on dominant species, and should overcome any problems posed by extant species that may have changed their preferred environment with time.

Interpretation of results

The ecologic range of species associations and faunal sample associations are outlined below. Full faunal lists, including suborder, age, and ecologic distribution, are given in Appendix 3c.

(a) Arawa-1 (Fig 6.7)

Inspection of the dendrogram classification of samples and species have resulted in the recognition of seven sample associations (left side of figure) and eleven distinct species associations (top of figure).

Species associations:

Aw1 – *Ammonia parkinsoniana* f. *aoteana* (restricted to brackish environments; Hayward et al., 1999), *Haynesina depressula* (inner to mid shelf, c.0-75 m; Hayward et al., 1999), *Elphidium charlottense* (inner to mid shelf, most abundant <20 m), and *Zeaflorilus parri* (mostly shallower than 25 m). Indicative of brackish environments to mid inner-shelf (<20 m).

Aw2 – *Elphidium excavatum* (dominantly brackish, but also found in the inner shelf), and *Gavelinopsis* sp. (inner and mid shelf). Inner shelf (<50 m).

Aw3 – *Quinqueloculina seminula* (abundant in shallow exposed environments; inner-shelf, and low numbers to mid shelf). Inner shelf (<50 m).

Aw4 –*Astrononion kickinskii* (probable brackish to inner shelf depths), *Bulimina elongata?* (inner shelf), and *Notorotalia zealandica* (dominantly inner shelf). Inner shelf (<50 m).

Aw5 – *Anomalinoides sphericus* (rare specimens; deep inner to mid shelf, also common at outer shelf and upper bathyal depths), *Oridorsalis tenera* (probable ecophenotype of *Oridorsalis umbonatus* that is common at outer shelf and bathyal depths, uncommon in shallower environments. May occur slightly shallower than *O. umbonatus*), and *Uvigerina* spp. (range from mid shelf to abyssal, but most common upper-mid bathyal). Outer shelf to mid bathyal depths based on predominance of *Uvigerina* (100-1000 m).

Aw6 – *Astrononion novozealandicum* (inner shelf to bathyal depths, most abundant outer shelf and bathyal), *Cassidulina carinata* (most abundant from 100-1500 m; Hayward et al., 2001), and *Notorotalia finlayi* (inner and mid shelf). Given dominance of *A. novozealandicum* and *C. carinata*, depth range estimated as mid to outer shelf, 50–600 m.

Aw7 – *Evolvocassidulina orientalis* (inner and mid shelf, though recorded most abundantly at upper bathyal depths), *Nonionellina flemingi* (abundant in deeper parts of quiet sounds (20-50 m) common at mid shelf to bathyal depths), and *Notorotalia* sp1. (genus *Notorotalia* displays greatest abundance at inner shelf and outer shelf to upper mid bathyal depths). Species in this association suggest that the depth range could either be inner to mid shelf, or outer shelf to upper bathyal depths. Given position in relation to sample associations (see below) it is more probable that this species association carries an inner shelf to mid shelf (20-100 m) signature.

Aw8 – *Anomalinoides parvumbilius* (outer shelf to upper bathyal?), *Bulimina striata* (common at bathyal depths), *Siphouvigerina* spp. (mid to lower bathyal?), *Hoeglundina elegans* (mid shelf to bathyal), and *Cibicides* sp4. (inner shelf to abyssal depths). Varied depth range identified from outer shelf to lower bathyal (100-1000 m).

Aw9 – *Bolivinita plioblіqua* (extinct; >200 m), *Haeuslerella parri* (extinct; >200 m ± 50 m at Taimana-1), *Notorotalia taranakia* (outer shelf – upper bathyal), and *Lenticulina* sp2. (greatest abundance of genus *Lenticulina* occurs at outer shelf to upper lower bathyal depths). Outer shelf to lower bathyal (200-1200 m) depths inferred.

Aw10 – *Bulimina marginata* f. *aculeata* (upper to lower bathyal), *Cibicides deliquatus* (outer shelf to mid bathyal), *Cibicides* sp3. (as for *Cibicides* sp4), and *Pullenia bulloides* (outer shelf to upper bathyal at shallowest). Outer shelf to mid bathyal (150-1000 m).

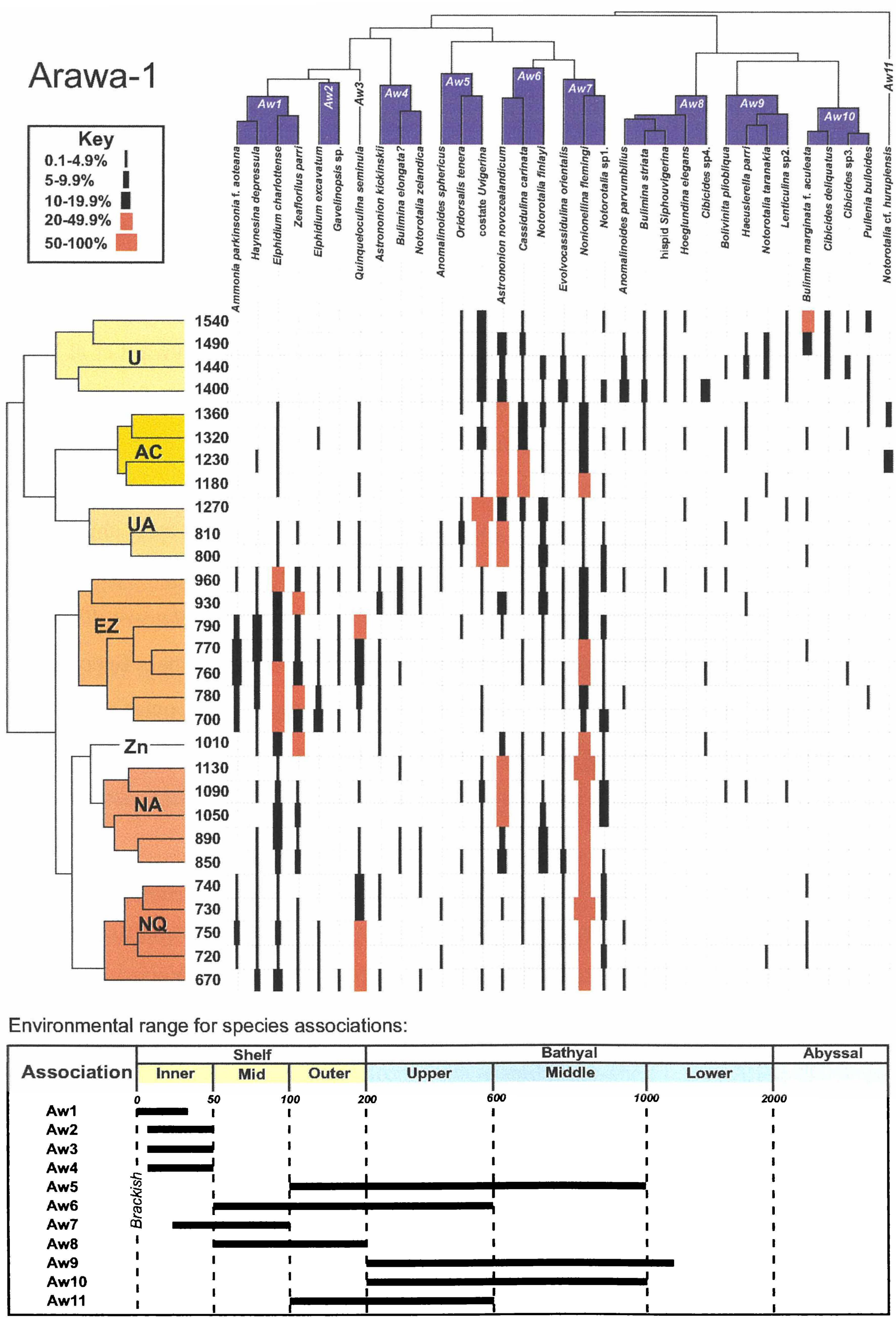


Fig. 6.7: Cluster analysis of benthic species, Arawa-1. Species associations (top, blue) are clustered using modified Morista Similarity, on species with more than 10 specimens. Faunal sample associations (left, shades of orange) are clustered using Bray-Curtis Distance Matrix, on samples with greater than 40 specimens.

Aw11 – *Notorotalia* cf. *hurupiensis* (extinct; most abundant at upper bathyal depths). ?Outer shelf to upper bathyal (100-600 m).

Faunal sample associations:

Association U (sample intervals –1400, 1440, 1490, and 1540 m)

Dominated by *Uvigerina* spp. (Aw5) but may be co-dominated by *Astrononion novozealandicum* (Aw6), *Bulimina marginata* f. *aculeata* or *Cibicides deliquatus* (Aw10). The presence of *Eggerella braydi*, *Sphaeroidina bulloides*, and *Cibicides neoperforatus* above 1490 m bKB, and *K. cylindrica*, *Stilostomella* sp. below 1490 m indicates deep outer shelf-upper slope depths, deepening to mid bathyal depths down-hole.

Association AC (sample intervals – 1180, 1230, 1320, and 1360 m)

Association dominated by *Astrononion novozealandicum*, and *Cassidulina carinata* of species association Aw7, with strong populations of *Nonionellina flemingi* (Aw8). Rare occurrences of *Pullenia bulloides*, *Hauslerella parri* and *Sphaeroidina bulloides* below 1300 m suggest shallowing up-hole from upper bathyal to outer shelf depths.

Association UA (sample intervals – 800, 810, and 1270 m)

Dominated by costate *Uvigerina* (Aw5; over 50% at 1270 m). Strong populations of *Astrononion novozealandicum* and *Notorotalia finlayi* (Aw6 and Aw7). Probable outer shelf to mid upper bathyal depths.

Association EZ (sample intervals – 700, 760, 770, 780, 790, 930, and 960 m)

Dominated by shallow water (mixed inner shelf and brackish) fauna (Aw1) particularly *Elphidium charlottense* and *Zeaflorilus parri*. Varying amounts *of Quinqueloculina seminula* (Aw3) and *Nonionellina flemingi* (Aw7). Occurrences of *Anomalinoides sphericus*, *Bolivinita pliobliqua*, and *Cassidulina carinata* below 900 m suggests mid to outer shelf depths for lower sample intervals, shallowing to inner shelf depths for upper sample intervals.

Association ZN (sample interval - 1010 m)

Co-dominance of *Zeaflorilus parri* (Aw1) and *Nonionellina flemingi* (Aw7) with strong populations of *Elphidium charlottense* (Aw1) strongly indicates inner shelf depths. However, the occurrence of *Bulimina marginata* f. *aculeata* and *Cassidulina carinata* suggests deeper (mid to outer shelf) depths.

Association NA (sample intervals – 850, 890, 1050, 1090, and 1130 m)

Very large populations of *Nonionellina flemingi* (Aw7) supported by strong populations of *Astrononion novozealandicum* (Aw6). Rare occurrences of *Haeuslerella parri* and *Bulimina marginata* f. *aculeata* indicate outer shelf to upper bathyal depths for some samples, while the presence of *Elphidium charlottense* suggests mixing of shallower water fauna. Mid shelf to upper bathyal depths interpreted.

Association NQ (sample intervals – 670, 720, 730, 740, and 750 m)

Co-dominance of *Nonionellina flemingi* (Aw7) and *Quinqueloculina seminula* (Aw3). Deeper inner shelf depths suggested, though sporadic occurrences of *Uvigerina*, *Anomalinoides sphericus*, and *Bulimina marginata* f. *aculeata* below 720 m suggest mid to outer shelf depths for deeper sample intervals.

Depths obtained in this study (refer to summary Fig. 6.8) are similar to, although somewhat more detailed than, those interpreted by Crundwell et al. (1992). Both studies identify *Karreriella cylindrica* at approximately the same interval (1440 m this study, 1435 m in the previous). This is consistent with the outer shelf to bathyal depths interpreted at this interval. Other bathyal-restricted species, such as *Siphonina australis* (1540 m), *Cibicides neoperforatus* (1490 m), and *Haeuslerella parri* (1360 m) indicate a progressive shallowing trend up-hole. General paucity of fauna below 1540 m has negated the use of samples below this depth. However, Crundwell et al. (1992) identify *Sigmoilopsis schlumbergeri* at 1578 m, which possibly marks submergence of the site into the middle bathyal zone. Planktic ratios support this, with oceanic water masses indicated below 1550 m. Middle bathyal conditions are interpreted throughout the Opoitian and Kapitean at this site, shallowing to upper bathyal during the Waipipian to Early Nukumaruan, and shelfal depths by the late Nukumaruan.

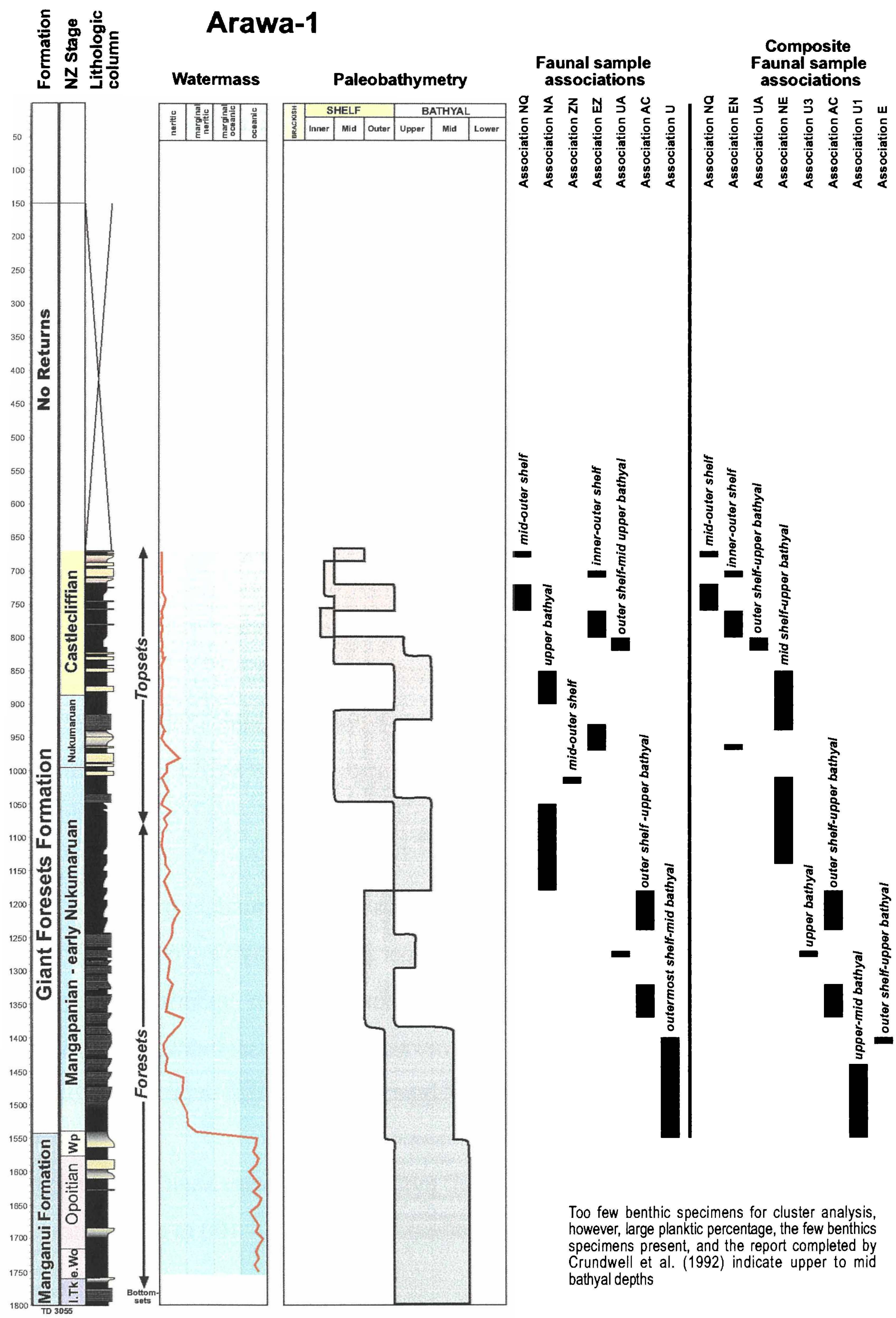


Fig. 6.8: Inferred paleobathymetry for Arawa-1. Depth ranges for faunal sample associations are discussed in text, and illustrated in Fig. 6.7. Composite faunal associations are illustrated in Fig. 6.15.

(b) Ariki-1 (Fig. 6.9)

Cluster analysis has resulted in dendrograms that have been divided into twelve species associations, and eight faunal sample associations. The overall signature is one of much greater depths than indicated by the associations observed at Arawa-1.

Species associations:

Ar1 – *Astrononion kickinskii* (mid shelf to lower bathyal based on other Astrononion species; Lewis, 1979), and *Cibicides deliquatus* (outer shelf to mid bathyal). Outer shelf to mid bathyal (100-1000 m) depths indicated.

Ar2 – *Astrononion novozealandicum* (inner shelf to abyssal depths, most abundant outer shelf and bathyal), *Notorotalia* sp1. (dominant genus; greatest abundance from inner shelf to upper mid bathyal depths), *Cassidulina carinata* (most abundant from 100-1500 m), and *Evolvocassidulina orientalis* (inner and mid shelf, though recorded most abundantly at upper bathyal depths). Suggestive of depths ranging from outer shelf to lower bathyal (100-1000 m).

Ar3 – *Sphaeroidina bulloides* (mid shelf to abyssal, most common outer shelf to lower bathyal). Given lack of other diagnostic species, a broad depth range of outer shelf to mid bathyal (100-1000 m) is inferred.

Ar4 – *Bolivinita quadrilatera* (deep outer shelf to abyssal, most abundant upper-mid bathyal), *Bulimina marginata* f. *aculeata* (upper bathyal to upper abyssal; Hayward et al., 2001), costate *Uvigerina* (range from mid shelf to abyssal, but most common upper-mid bathyal), and *Oridorsalis tenera* (note comments made in Arawa-1; common at outer shelf and bathyal depths, uncommon in shallower environments). Upper to mid bathyal depths (200-1000 m).

Ar5 – *Gyroidina* sp1. (deep mid shelf to abyssal, most abundant outer shelf to upper bathyal), and *Karreriella cylindrica* (upper depth limit calibrated at 500 m +/- 100 m at this site). Upper bathyal (200-600 m).

Ar6 – *Cibicides* sp4. (inner shelf to abyssal depths), and *Pullenia bulloides* (deeper than mid outer shelf). ?Outer shelf to mid bathyal (250-1000 m).

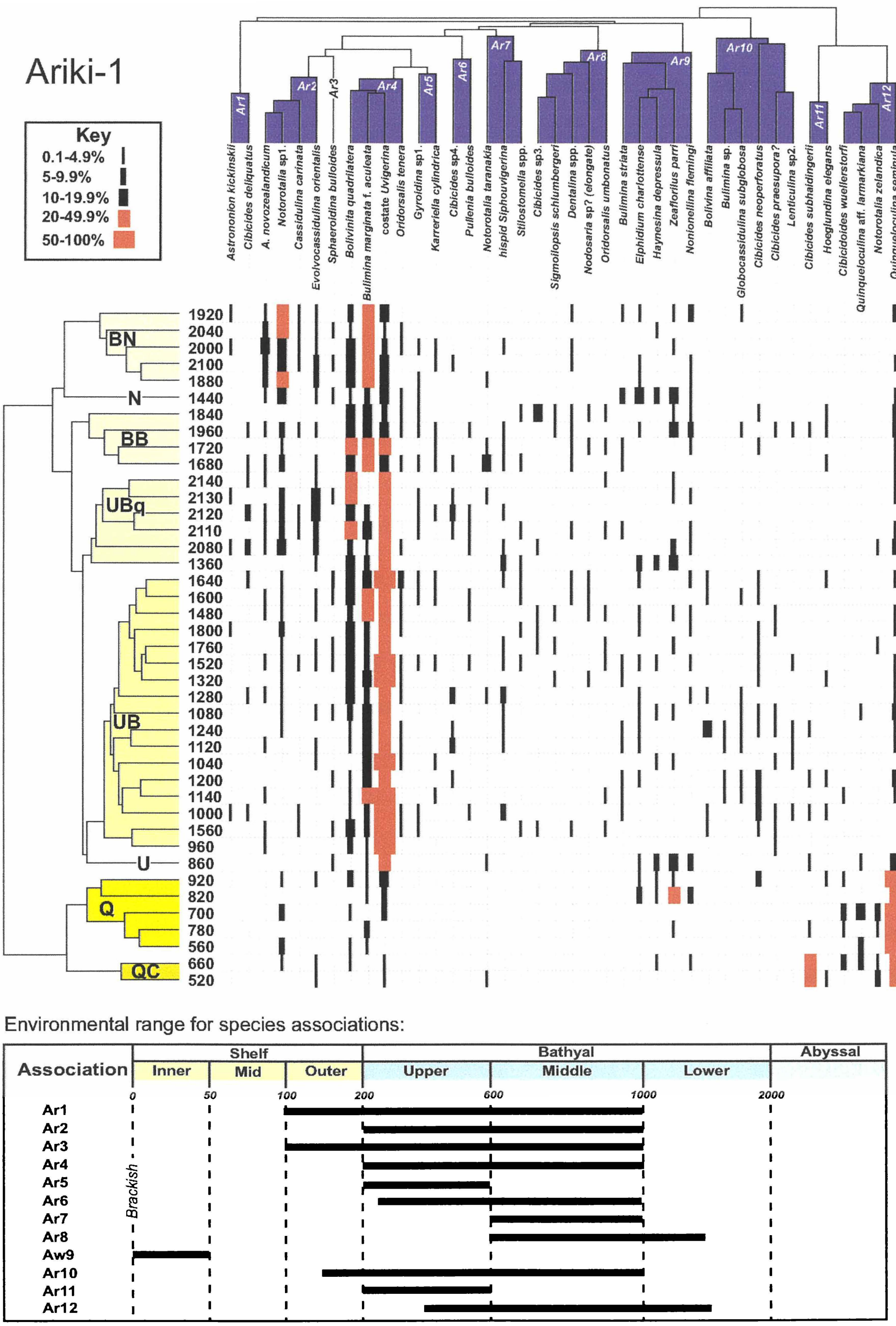


Fig. 6.9: Cluster analysis of benthic species, Ariki-1. Species associations (top, blue) are clustered using modified Morista Similarity, on species with more than 10 specimens. Faunal sample associations (left, shades of yellow) are clustered using Bray-Curtis Distance Matrix, on samples with greater than 40 specimens.

Ar7 – *Notorotalia taranakia* (outer shelf to ?mid bathyal), *Siphouvigerina* (deep outer shelf to ?mid bathyal), and *Stilostomella* spp. (upper depth limit c.500-600 m; B. Hayward, pers. comm., 2001). Mid bathyal depth range (600-1000 m).

Ar8 – *Cibicides* sp3. (as for *C.* sp4. above), *Sigmoilopsis schlumbergeri* (c. 600 m, to abyssal), *Dentalina* spp. (most common upper and mid bathyal), elongate *Nodosaria* (inner shelf to abyssal, but most abundant at depth greater than upper bathyal), and *Oridorsalis umbonatus* (most abundant at upper to lower bathyal depths). Presence of *S. schlumbergeri* caps upper depth range at c.600 m. Mid bathyal to mid lower bathyal (c.600-1500 m).

Ar9 – *Bulimina striata* (common at bathyal depths) *Elphidium charlottense* (inner to mid shelf, most abundant <20 m), *Haynesina depressula* (inner to mid shelf, c.0-75 m), *Zeaflorilus parri* (mostly shallower than 25 m), and *Nonionellina flemingi* (abundant in deeper parts of quiet sounds (20-50 m)). Inner shelf (<20 m).

Ar10 – *Bolivinita affiliata* (extinct; most abundant at bathyal depths), *Globocassidulina subglobosa* (most abundant at bathyal depths, with upper depth limit of mid outer shelf), *Cibicides neoperforatus* (upper to lower bathyal), *Cibicides praesupora*? (outer shelf?), and *Lenticulina* sp2. (greatest abundance of genus *Lenticulina* occurs at outer shelf to upper bathyal depths; Hayward, 1986). Large depth range from outer shelf to upper lower bathyal (c.150-1000 m)

Ar11 – *Cibicides subhaidingerii* (outer shelf to ?mid bathyal), and *Hoeglundina elegans* (widespread at mid shelf to bathyal depths; >200 m). Deep outer shelf to mid bathyal (c.200-600 m).

Ar12 – *Cibicidoides wuellerstorfi* (>500 m ± 150m; possibly mis-identified), *Quinqueloculina* aff. *lamarkiana* (mostly inner shelf), *Notorotalia zealandica* (dominantly inner shelf), and *Quinqueloculina seminula* (abundant in shallow exposed environments, inner-shelf, and low numbers to mid shelf). Seismic position and facies (degradational) suggest basinward transport of inner shelf facies to an upper bathyal position (c.400-1500 m).

Faunal sample associations:

Association BN (sample intervals – 1880, 1920, 2000, 2040, and 2100 m)

Equally dominated by *Bulimina marginata* f. *aculeata* (Ar 4), and *Notorotalia* sp1. (Ar2), with moderate populations of *Astrononion novozealandicum* (Ar2) *Bolivinita quadrilatera* and *Uvigerina* spp. (Ar4). Upper to mid bathyal depths.

Association N (sample interval – 1440 m)

Moderate populations of *Notorotalia* sp1. (Ar2), but no real dominants. Reasonable contribution from shallower water fauna of species association Ar9, including *Elphidium charlottense*, *Haynesina depressula*, and *Zeaflorilus parri.* Depositional environment interpreted as upper bathyal depths, with considerable transportation and mixing of inner shelf fauna.

Association BBq (sample intervals – 1680, 1720, 1840, and 1960 m)

Dominated by *Bulimina marginata* f. *aculeata* and *Bolivinita quadrilatera*, with considerable contributions from costate *Uvigerina* (all from species association Ar4). Upper to mid bathyal depths. Presence of *Sigmoilopsis schlumbergeri* at 1840 and 1960 m, and *Stilostomella* at 1680 and 1840 m may restrict these samples to mid bathyal depths or greater (c. 600 and deeper).

Association UBq (sample intervals – 1360, 2080, 2110, 2120, 2130, and 2140 m)

Dominated by costate *Uvigerina*, with large populations of *Bolivinita quadrilatera* (Ar4). Moderate populations of *Notorotalia* sp1. and *Evolvocassidulina orientalis* (Ar2), and sporadic occurrence of *Stilostomella* sp. suggests depths are restricted to lower-most upper to mid bathyal (c. 500-1000 m). Predominance of deeper water faunas, with little contribution from shallower water faunas, in the deeper samples, suggests sediment from inner-mid shelf regions is not being transported to this site during this time (Opoitian – only very thin Opoitian recorded).

Association UB (sample intervals – 960, 1000, 1040, 1080, 1120, 1140, 1200, 1240, 1280, 1320, 1480, 1520, 1560, 1600, 1640, 1760, and 1800 m)

Very large association, clearly dominated by *Uvigerina* spp. (always greater than 20%, and often over 50% of faunal assemblage), with varying contributions from (particularly) *Bulimina marginata* f. *aculeata* and *Bolivinita quadrilatera* (Ar4). Upper to mid bathyal depths. Sample intervals 1080, 1320 (and 1300 m; Hayward, 1986), 1480, and 1760 m include specimens of

Sigmoilopsis schlumbergeri, restricting upper depth limits to approximately 600 m for these sample intervals. The presence of *Stilostomella* spp. below 1500 m supports an upper depth limit of c. >500-600 m.

Association U (sample interval – 860 m)

Dominated by *Uvigerina* (Ar4), but with significant contributions of *Zeaflorilus parri* (Ar9), and *Quinqueloculina seminula* (Ar12). Suggestive of deep outer shelf to upper bathyal depths with transport and mixing of shallower water faunas.

Association Q (sample intervals – 560, 700, 780, 820, and 920 m)

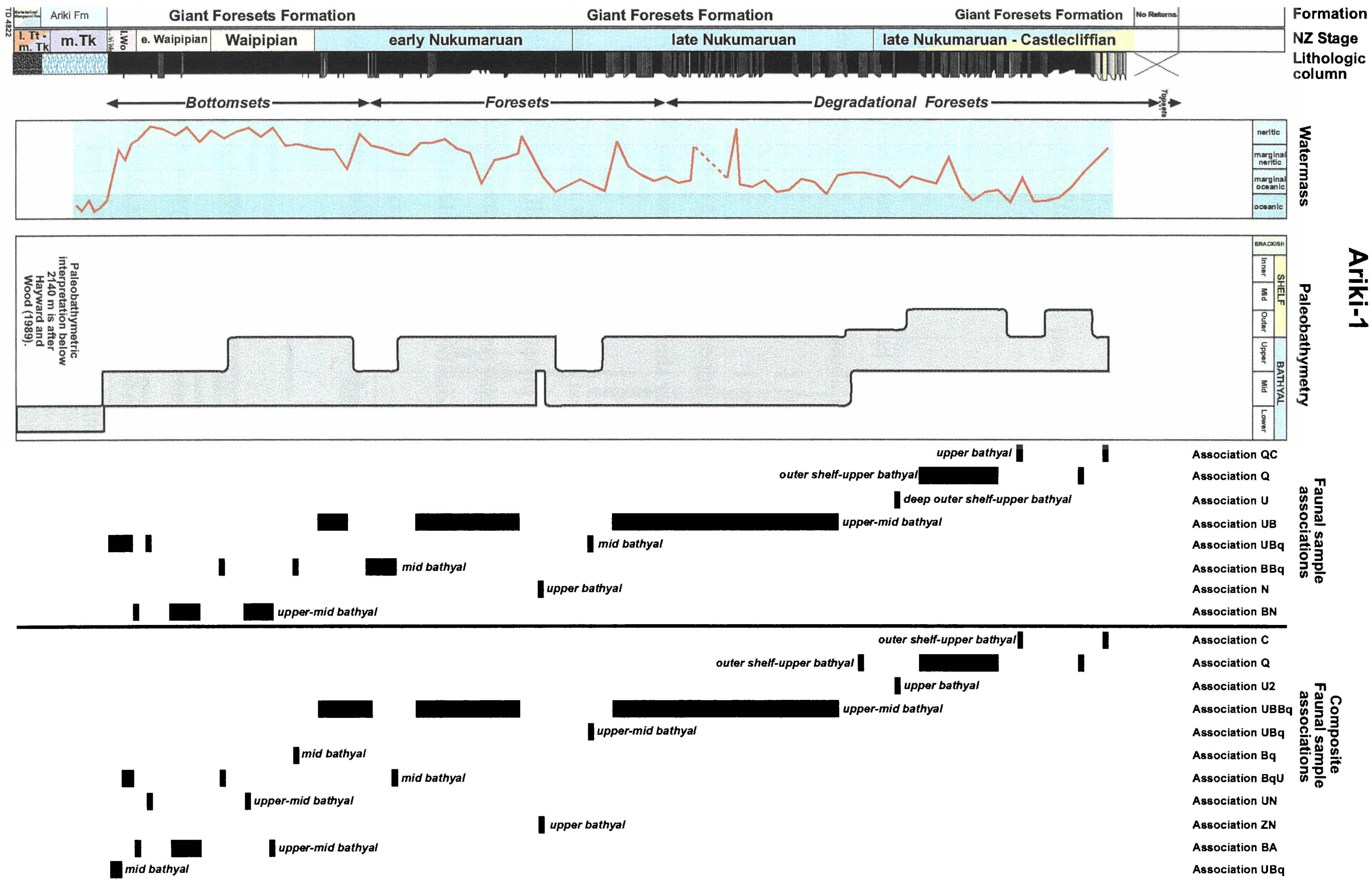
Dominated by *Quinqueloculina seminula* (Ar12; mostly greater than 50% of faunal association). Inner shelf depths initially indicated for majority of samples, but presence of deeper water faunas at 920 m (e.g., *Hoeglundina elegans*, *Cibicides neoperforatus, Cibicides wuellerstorfi*), and *Cibicides subhaidingerii* at 820 and 780 m, suggest deeper outer shelf to upper bathyal depths. Considerable sediment remobilisation and transport from inner shelf depths and/or lower sea level is inferred.

Association QC (sample intervals – 520 and 660 m)

Co-dominated by *Quinqueloculina seminula* (Ar12), and *Cibicides subhaidingerii* (Ar11). Presence of ?*Cibicides wuellerstorfi* and *Bolivinita quadrilatera* indicates upper bathyal depths, with some transport and mixing of shallow-water (inner shelf) fauna.

Figure 6.10 shows a steady decrease in water depth through the Plio-Pleistocene, after an initial and apparently sharp decrease in water depth somewhere between middle Kapitean and Late Opoitian. Depths remain in the upper to mid bathyal range throughout the Late Opoitian to Late Nukumaruan, with possible intermittent shallowing and deepening trends. However, depth ranges are too broad to be able to confidently suggest that these slight perturbations are related to global sea level fluctuations. Both this study, and that of Hayward (1986b) noted considerable quantities of inner shelf faunas at 1440 m, (also at 1360 m, this study). Hayward (1986b) suggests that this may indicate periods of lowered sea level (with reworking of fauna during glacials) though inherent downhole contamination of cuttings samples may be partially responsible for the mixing of faunas.

Fig. 6.10 (facing page): Inferred paleobathymetry for Ariki-1. Depth ranges for faunal sample associations are discussed in text, and illustrated in Fig. 6.9. Composite faunal associations are illustrated in Fig. 6.15.


Ariki-1
Formation
NZ Stage
Lithologic column
Ariki Fm
Giant Foresets Formation
Giant Foresets Formation
Giant Foresets Formation
No Returns
TD 4822
l. Tt - m. Tk
m.Tk
l.Wo
e. Waipipian
Waipipian
early Nukumaruan
late Nukumaruan
late Nukumaruan - Castlecliffian
Bottomsets
Foresets
Degradational Foresets
Topsets
Watermass
neritic
marginal neritic
marginal oceanic
oceanic
Paleobathymetry
BRACKISH
SHELF
Inner
Mid
Outer
BATHYAL
Upper
Mid
Lower
Paleobathymetric interpretation below 2140 m is after Hayward and Wood (1989).
Faunal sample associations
upper bathyal
Association QC
outer shelf-upper bathyal
Association Q
deep outer shelf-upper bathyal
Association U
upper-mid bathyal
Association UB
mid bathyal
Association UBq
mid bathyal
Association BBq
upper bathyal
Association N
upper-mid bathyal
Association BN
Composite faunal sample associations
outer shelf-upper bathyal
Association C
outer shelf-upper bathyal
Association Q
upper bathyal
Association U2
upper-mid bathyal
Association UBBq
upper-mid bathyal
Association UBq
mid bathyal
Association Bq
mid bathyal
Association BqU
upper-mid bathyal
Association UN
upper bathyal
Association ZN
upper-mid bathyal
Association BA
mid bathyal
Association UBq

More distinct sea level changes are noted in the Late Nukumaruan to Castlecliffian. Cuttings below 2140 m had too few benthic foraminifera to include in cluster analysis, but those that are present indicate consistently deeper depths (mid bathyal – 1000-1500 m), and show no evidence of imported shallow water taxa. Because cuttings sample 2150 m crosses the boundary between Wanganui and Taranaki Series strata, Hayward (1986b) suggests that deeper faunas are derived from the older sediments.

(c) Kora-1 (Fig. 6.11)

Kora-1 samples yielded eleven species associations and nine faunal sample associations on examination of the cluster dendrograms.

Species associations:

K1 – *Astrononion kickinskii* (inner shelf to brackish, but ranges through to mid bathyal), and *Elphidium charlottense* (inner to mid shelf, most abundant <20 m). Brackish to inner shelf environment indicated (<20 m).

K2 – *Cibicides subhaidingerii* (outer shelf to mid upper bathyal). As there is only one species in this association, range can only be given as outer shelf to mid upper bathyal (100-400 m).

K3 – *Nonionellina flemingi* (mid shelf to bathyal, and deep, quiet inlets), *Quinqueloculina seminula* (inner shelf, <50 m), *Notorotalia finlayi* (inner and mid shelf, also abundant in deeper and quieter parts of enclosed inlets), and *Notorotalia zelandica* (predominantly inner shelf, but ranges to mid shelf). Inner shelf environments, possibly enclosed inlet (<50 m).

K4 – *Notorotalia depressa* (inner shelf), and *Zeaflorilus parri* (mostly shallower than 25 m). Inner shelf (<25 m).

K5 – *Astrononion novozealandicum* (inner and mid shelf, more abundant at outer shelf and upper bathyal depths), *Bulimina marginata* f. *aculeata* (upper bathyal to abyssal) *Cassidulina carinata* (most abundant from 100-1500 m), and *Evolvocassidulina orientalis* (inner and mid shelf, most abundant at upper bathyal depths). Indicative mid shelf to upper bathyal environments (c.50-600 m).

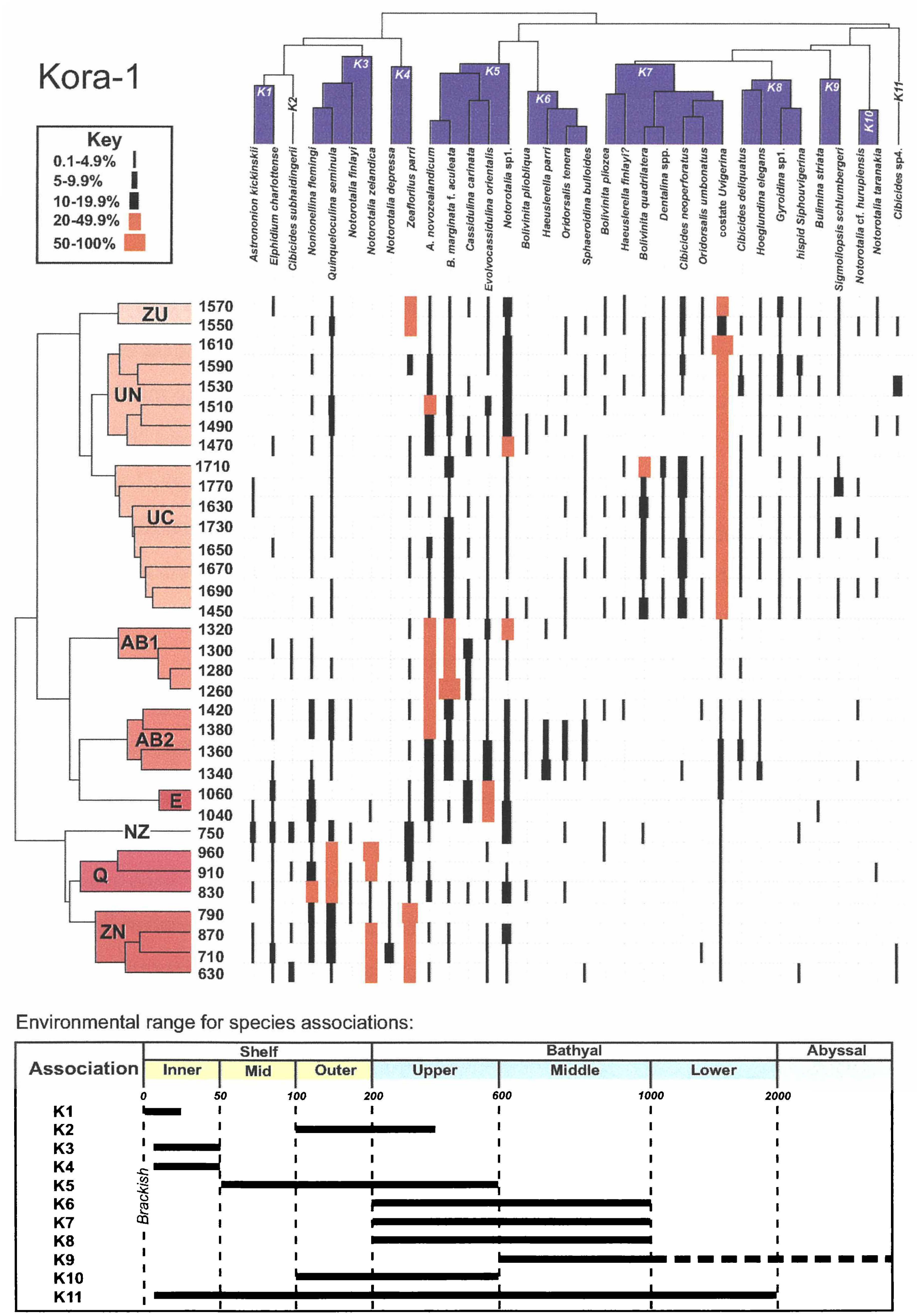


Fig. 6.11: Cluster analysis of benthic species, Kora-1. Species associations (top, blue) are clustered using modified Morista Similarity, on species with more than 10 specimens. Faunal sample associations (left, shades of red) are clustered using Bray-Curtis Distance Matrix, on samples with greater than 40 specimens.

K6 – *Bolivinita pliobliqua* (extinct; >200 m) *Haeuslerella parri* (extinct; upper bathyal), *Oridorsalis tenera* (mid shelf and bathyal), and *Sphaeroidina bulloides* (mid shelf to abyssal, most abundant at outer shelf to upper bathyal depths). Upper to mid bathyal (c.200-1000 m) depths.

K7 – *Bolivinita pliozea* (extinct; most abundant at upper bathyal depths), *Haeuslerella finlayi* (mid shelf to mid bathyal), *Bolivinita quadrilatera* (deep outer shelf to abyssal, most abundant at upper to mid bathyal depths), *Dentalina* spp. (most common upper to mid bathyal), *Cibicides neoperforatus* (upper to lower bathyal) *Oridorsalis umbonatus* (most abundant upper to lower bathyal) and *Uvigerina* spp. (most abundant upper to mid bathyal). Upper to mid bathyal (200-1000 m).

K8 – *Cibicides deliquatus* (outer shelf to mid bathyal), *Hoeglundina elegans* (widespread at mid shelf to bathyal depths; ~200 m at Ariki-1), *Gyroidina* sp. (deep mid shelf to abyssal, most common outer shelf to upper bathyal), and *Siphouvigerina* spp. (mid to lower bathyal). Wide environmental range suggested, from upper to mid bathyal (200-1000 m).

K9 – *Bulimina striata* (common at bathyal depths), and *Sigmoilopsis schlumbergeri* (upper mid bathyal to abyssal, c.600 m upper depth limit). Interpreted as no shallower than upper mid bathyal (>c.600 m).

K10 – *Notorotalia* cf. *hurupiensis* (outer shelf to upper bathyal), and *Notorotalia taranakia* (outer shelf to upper bathyal). Outer shelf to upper bathyal depths (c.100-600 m) based on similar environmental ranges.

K11 – *Cibicides* sp4. (inner shelf to abyssal). Wide range of genus *Cibicides* (refer Figure 6.5) results in a large environmental range, from inner shelf to lower bathyal (<50-2000 m).

Faunal sample associations:

Association ZU (sample intervals – 1550 and 1570 m)

Dominated by *Zeaflorilus parri* (K4), and *Uvigerina* spp. (K7), with good populations of *Notorotalia* sp1. (K5). The presence of *Hoeglundina elegans* and (particularly) *Sigmoilopsis schlumbergeri* in both these samples, and *Cibicides wuellerstorfi* at 1570 m, results in an uppermost mid bathyal depth range (c. 600 m), with considerable input from transported shallow water-derived fauna. Mid bathyal depths indicated.

Association UN (sample intervals – 1470, 1490, 1510, 1530, 1590, and 1610 m)

Dominated by *Uvigerina* (K7), but unlike previous association, has a noticeable lack of *Zeaflorilus parri*. Instead, *Notorotalia* sp1. and *Astrononion novozealandicum* appear in greater numbers (K5), with varying contributions from *Hoeglundina elegans*, *Cibicides neoperforatus* and *Haeuslerella* spp. Upper bathyal depths indicated, becoming deeper (mid bathyal) down-hole (*Sigmoilopsis schlumbergeri* noted at 1590 and 1610 m).

Association UC (sample intervals – 1450, 1630, 1650, 1670, 1690, 1710, 1730, and 1770 m)

Dominated by *Uvigerina* spp., with strong populations of *Cibicides neoperforatus*, as well as *Bolivinita quadrilatera* (both K7), and also *Bulimina marginata* f. *aculeata* (K5). *Sigmoilopsis schlumbergeri* occurs in good numbers in all samples except for 1630-1670 m. Mid bathyal depths interpreted.

Association AB1 (sample intervals – 1260, 1280, 1300, and 1320 m)

The first of two faunal sample associations that are dominated both by *Astrononion novozealandicum* and *Bulimina marginata* f. *aculeata* (K5). Only small populations of a limited number of subsidiary species, including *Cassidulina carinata* (<20%, and generally less than 10%) and *Evolvocassidulina orientalis* (<10%) are present. Presence of *H. parri* at 120 m bKB indicates deepening down-hole. Interpreted as outermost shelf to upper bathyal depositional environments.

Association AB2 (sample intervals – 1340, 1360, 1380, and 1420 m)

As above, but has a greater number of subsidiary species, including *Cassidulina carinata*, *Evolvocassidulina orientalis*, and *Notorotalia* sp1, as well as minor (<10%) populations of *Uvigerina* spp. (K7). Dominant species occur in lower percentages than seen in AB1. The occurrence of deeper water taxa (e.g., *Hoeglundina elegans* and *Haeuslerella* spp. in most sample intervals, and *Stilostomella* sp. at 1420 m), suggest an overall deepening environment from the previous association (to mid bathyal).

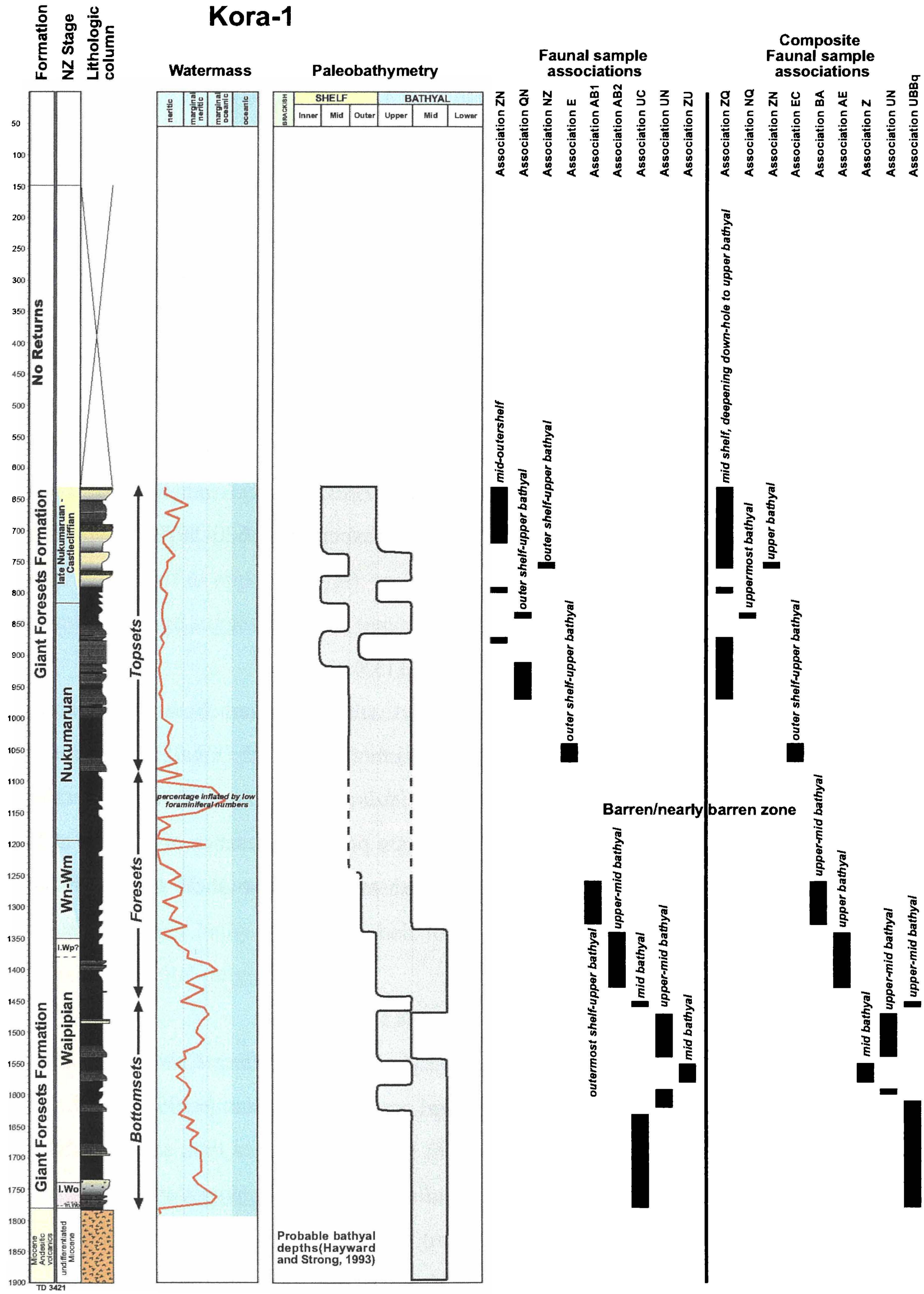


Fig. 6.12: Inferred paleobathymetry for Kora-1. Depth ranges for faunal sample associations are discussed in text, and illustrated in Fig. 6.11. Composite faunal associations are illustrated in Fig. 6.15.

Association E (sample intervals 1040 and 1060 m)

Dominated by *Evolvocassidulina orientalis*, with strong populations *of Astrononion novozealandicum*, *Cassidulina carinata*, and *Notorotalia* sp1. (all K5). Presence of *Bulimina marginata* f. *aculeata* suggests outer shelf to upper bathyal depths.

Association NZ (sample interval – 750 m)

Hugely diverse sample with real dominant components, although *Notorotalia* sp1. (K5) and *Zeaflorilus parri* (K4) comprise 10-19.9% of samples. Other minor species include shallower water faunas from species association K1, and K3, with the slightly deeper K2 benthic, *Cibicides subhaidingerii*. The diversity of fauna from these associations, and the presence of other deeper water fauna such as *Globocassidulina subglobosa* and *Bolivinita quadrilatera*, suggests transport from inner shelf depths, and mixing at outer shelf to upper bathyal depths.

Association QN (sample intervals –830, 910, and 960 m)

Dominated by *Quinqueloculina seminula* (K3) with some large populations of *Notorotalia zealandica* and *Nonionellina flemingi* (also K3). The occurrence of deeper water species (e.g., *Cibicides subhaidingerii*, *Astrononion novozealandicum*, and *Bulimina marginata* f. *aculeata*) indicates mid to shelf to uppermost bathyal depths.

Association ZN (sample intervals – 630, 710, 790, and 870 m)

Large populations of *Zeaflorilus parri* (K4; 20-49.9%, and over 50% at 790 m) and *Notorotalia zelandica*, with variable populations of both *Nonionellina flemingi* and *Quinqueloculina seminula* (K3) occur together with *Cibicides subhaidingerii* (K2) and *Evolvocassidulina orientalis* (K5). Sporadic occurrences of deeper water species (e.g., *Astrononion novozealandicum*, *Sphaeroidina bulloides*, and *B. marginata* f. *aculeata*) indicate mid to outer shelf depths.

Paleodepths change from dominantly mid to upper bathyal depths during the Late Opoitian and Waipipian, to dominantly mid to outer shelf depths by the Nukumaruan (Fig. 6.12), coinciding with a change in sedimentation from predominantly muddy below 1000 m, to more sand rich above 1000 m. This suggests that sea level lowered enough to allow sediment to be transported further across the shelf, and be deposited at this site, or that sea level was low enough that shallower-water taxa could become established. Most faunas above 1000 m have a distinct lack of, or very few, deeper water taxa, suggesting a combination of the above factors. While

Hayward and Strong (1993) record the first occurrence of a bathyal restricted taxa at 1700 m bKB, this study first records *Cibicides neoperforatus* and *Haeuslerella parri* at 1340 m. These taxa are consistently present through to 1590 m. *Sigmoilopsis schlumbergeri* is first noted at 1450 m, and is present in most samples below this depth. This indicates that upper depth limits below 1340 m are restricted to ~ 200 m (uppermost bathyal), while below 1450 m many samples have an upper depth limit of c.600 m. The occurrence of shallower water faunas is attributed to remobilisation and transport of inner shelf sediments. Depths below 1770 m (within volcanics) are non-fossiliferous, but it is thought that the volcanic pile was submarine and built up at bathyal depths during the middle to Late Miocene.

(d) Wainui-1 (Fig. 6.13)

Thirteen species associations and eight faunal sample associations have been identified from dendrograms generated for Wainui-1. While all wells tend to have a dominant species that is found in many of the sample associations, *all* faunal sample associations for Wainui-1, bar one, are dominated by costate *Uvigerina*. The one grouping in which *Uvigerina* do not dominate, is instead dominated by hispid *Siphouvigerina* (which belongs to the same subfamily; Loeblich and Tappan, 1987). This dominance by one species presumably reflects the more distal position that Wainui-1 held through the Pliocene and Pleistocene, and hence possibly it's heightened sensitivity to changing current systems (see section 6.4).

Species associations:

W1 – *Anomalinoides parvumbilius* (most abundant upper bathyal), *Bolivinita pliozea* (extinct; most abundant at upper bathyal depths), *Sphaeroidina bulloides* (mid shelf to abyssal, most abundant outer shelf-upper bathyal), *Bulimina striata* (common at bathyal depths), and hispid *Siphouvigerina* (dominant mid to lower bathyal). Upper to mid bathyal depths (200-1000 m).

W2 – *Karreriella cylindrica* (depth calibrated at this site to >200 m +/– 50 m; Hayward, 1990). Upper bathyal and deeper (> c.200 m).

W3 – *Notorotalia taranakia* (mid shelf-upper bathyal). Mid shelf to upper bathyal depths (c.100-600 m).

W4 – *Bolivinita quadrilatera* (deep outer shelf-abyssal, most abundant upper-mid bathyal), *Dentalina* spp. (most common upper-mid bathyal), *Oridorsalis tenera* (mid shelf-bathyal), costate *Uvigerina* (most abundant upper to mid bathyal), *Bulimina marginata* f. *aculeata* (upper bathyal to abyssal), *Stilostomella* spp. (mid bathyal - abyssal), and *Gyroidina* sp1. (deep mid shelf-abyssal, most abundant outer shelf-upper bathyal). Mid bathyal to abyssal (c.500->1000 m).

W5 – *Astrononion kickinskii* (brackish and inner shelf, ranges through to mid bathyal), *Cibicides* sp3. (most abundant inner shelf, and outer shelf-abyssal, rare mid shelf), *Discorbinella bertheloti* (deep inner shelf-bathyal, common at mid shelf), *Gavelinopsis hamatus* (inner-mid shelf), *Notorotalia zelandica* (inner-mid shelf, but especially inner shelf), and *Sigmoidella kagaensis*? (inner-mid shelf, and outer parts of harbours). All species strongly point to a deep inner to mid shelf environment (30-100 m).

W6 – *Astrononion novozealandicum* (inner and mid shelf, more abundant at outer shelf and bathyal depths), and *Nonionellina flemingi* (mid shelf to bathyal depths, and deeper quieter parts of inlets). Outer shelf to upper bathyal environment interpreted (100-400 m).

W7 – *Cassidulina carinata* (most abundant from 100-1500 m), *Evolvocassidulina orientalis* (inner and mid shelf, most abundant at upper bathyal depths), and *Notorotalia* sp1. (inner shelf to mid bathyal). Inner to mid bathyal depths (c.100->600 m).

W8 – *Elphidium charlottense* (inner to mid shelf, most abundant <20 m), *Zeaflorilus parri* (mostly shallower than 25 m), *Haynesina depressula* (inner to mid shelf, c.0-75 m), and *Quinqueloculina seminula* (inner shelf, low numbers out to mid shelf). Shallow inner shelf (<30 m) interpreted.

W9 – *Cibicides deliquatus* (outer shelf to mid bathyal), and *Siphotextularia wairoana* (inner-outer shelf?). Outer shelf to mid bathyal depths (100-1000 m).

W10 – Occurrence of only one species in this association (*Pyrgo anomala*), restricts the environment to its range of inner and mid shelf depths (<100 m).

W11 – *Cibicides* sp4. As genus *Cibicides* has a large environmental range (inner shelf to abyssal), interpretation must therefore follow this. Inner shelf - abyssal (c.0 - 5000 m) depths.

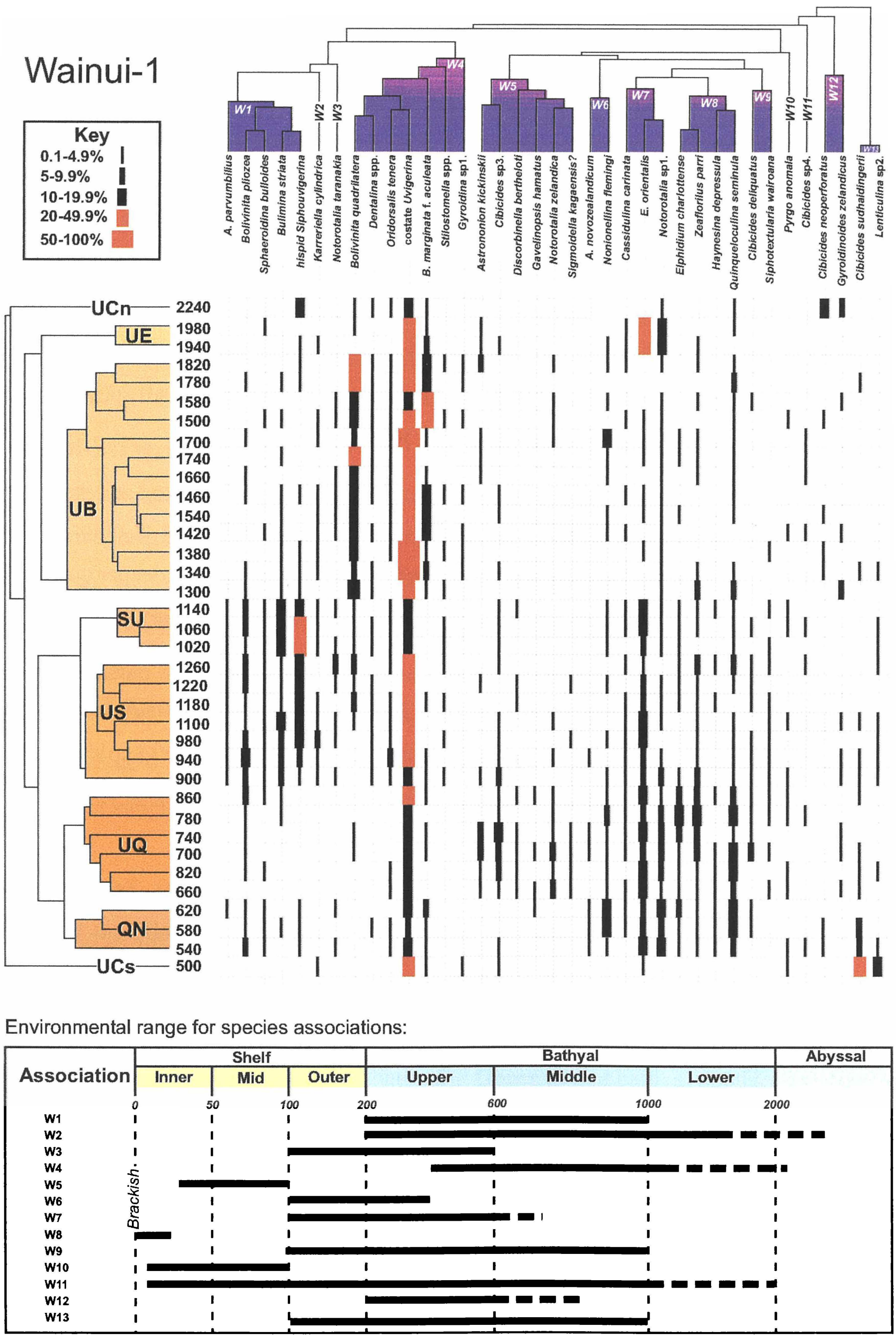


Fig. 6.13: Cluster analysis of benthic species, Wainui-1. Species associations (top, blue) are clustered using modified Morista Similarity, on species with more than 10 specimens. Faunal sample associations (left, shades of brown) are clustered using Bray-Curtis Distance Matrix, on samples with greater than 40 specimens.

W12 – *Cibicides neoperforatus* (upper-lower bathyal), and *Gyroidinoides zelandica* (most common at outer shelf to upper slope depths). Upper bathyal and deeper (>200 m).

W13 – *Cibicides subhaidingerii* (outer shelf-mid upper bathyal), and *Lenticulina* sp2. (dominant outer shelf-upper lower bathyal, low numbers inner-mid shelf and lower bathyal-abyssal). Outer shelf to mid upper bathyal depths (100-400 m).

<u>Faunal sample associations:</u>

Association UCn (sample interval – 2240 m)

No real dominants, though sample composed of 10-19.9% of costate *Uvigerina* (W4), *Cibicides neoperforatus* (W12), and also hispid *Siphouvigerina* (W1). Although the sample is dominated by species from a range of species associations, all point towards depths of mid bathyal and greater (supported by occurrence of *Sigmoilopsis schlumbergeri* and *Eggerella bradyi*).

Association UE (sample intervals – 1940 and 1980 m)

Clearly co-dominated by *Uvigerina* (W4), and *Evolvocassidulina orientalis* (W7; greater than 20% of each), with strong populations of *Notorotalia* sp1. (W7). Probable depositional environment is one of upper and mid bathyal depths, with transport and mixing of inner to mid shelf faunas.

Association UB (sample intervals – 1300, 1340, 1380, 1420, 1460, 1500, 1540, 1580, 1660, 1700, 1740, 1780, and 1820 m)

A large faunal association, strongly dominated by costate *Uvigerina* (>20%, and at times >50%), with large populations of *Bolivinita quadrilatera*, and smaller, though considerable, contributions from *Bulimina marginata* f. *aculeata* (all species association W4). Presence of *Sigmoilopsis schlumbergeri* in deeper sampled intervals (1660, 1780, and 1820 m bKB) and *Stilostomella* at 1300, 1380, 1460, 1500 and 1820 m suggests increasing paleodepths (>600 m) down-hole. Mid bathyal depths and greater interpreted.

Association SU (sample intervals – 1020, 1060, and 1140 m)

Siphouvigerina (W1; generally >20%), dominate over *Uvigerina* (W4; <20%), in this association, with good populations of *Bulimina striata*, *Bolivinita pliozea* (W1), and *Evolvocassidulina orientalis* (W7) also present. Upper to mid bathyal depths (>200 m).

Association US (sample intervals – 900, 940, 980, 1100, 1180, 1220, and 1260 m)

Dominated by costate *Uvigerina*, though not as strongly as seen in association UB. Hispid *Siphouvigerina* contribute markedly, with good numbers of other faunas from species association W1 (as above), and *Evolvocassidulina orientalis* from species association W7. The presence of *Stilostomella* spp. below 900 m suggests an upper depth limit of c.500 m. A mid to ?lower bathyal depositional environment is interpreted, with some transport and mixing of shallower water fauna.

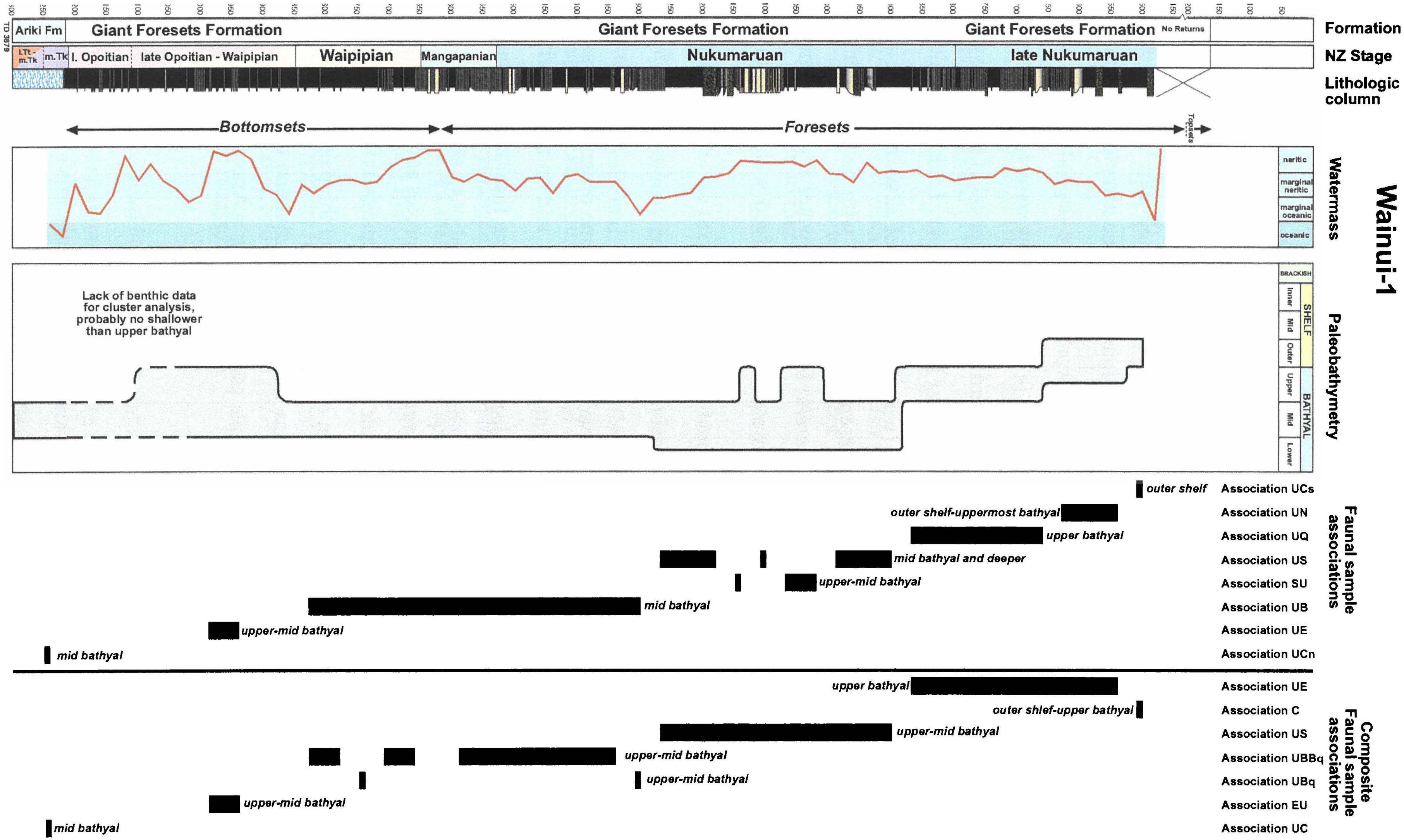
Association UQ (sample intervals – 660, 700, 740, 780, 820, and 860 m)

While this association is still dominated by costate *Uvigerina*, they do not occur in the same numbers as observed in previous associations (only 10-19.9%). *Quinqueloculina seminula* (W8), occur persistently and in reasonable numbers (often >10%), along with other shallower water fauna from that association (e.g., *Elphidium charlottense*, *Zeaflorilus parri*), and good populations of *Evolvocassidulina orientalis* (W7). The presence of numerous deeper water species, such as *Cibicides subhaidingerii, Bulimina marginata* f. *aculeata*, *Astrononion novozealandicum*, and *Bolivinita quadrilatera* indicate depths are probably still bathyal (upper). Though there are good numbers of inner shelf faunas, it is not considered that the site was as shallow as this at any time, and instead, transport across the shelf is inferred.

Association UN (sample intervals – 540, 580, and 620 m)

Similar to the previous association, with lower numbers of *Uvigerina* (<20%), and good contributions from a number of other species, including *Notorotalia* sp1., *Evolvocassidulina orientalis* (W7), *Nonionellina flemingi* (W6), and *Quinqueloculina seminula* (W8). Again, while this site is not considered to have been at inner shelf depths at any stage, depths as shallow as mid shelf, and possibly deep inner shelf, may have been obtained during lowered sea levels, although presence of *Cibicides subhaidingerii*, and *Cibicides neoperforatus* (540-580 m bKB) suggest upper bathyal depths. Seismic interpretation has indicated that sediment supply during the Pleistocene was voluminous, even at this more distal site, and it is feasible that large amounts of shelf-derived sediment were being carried as far as Wainui-1, especially during lowered sea levels.

Fig. 6.14 (facing page): Inferred paleobathymetry for Wainui-1. Depth ranges for faunal sample associations are discussed in text, and illustrated in Fig. 6.13. Composite faunal associations are illustrated in Fig. 6.15.



Wainui-1
Formation
NZ Stage
Lithologic column
Watermass
Paleobathymetry
Faunal sample associations
Composite Faunal sample associations
TD 3879
Ariki Fm
Giant Foresets Formation
Giant Foresets Formation
Giant Foresets Formation
No Returns
m.Tk
l. Opoitian
late Opoitian - Waipipian
Waipipian
Mangapanian
Nukumaruan
late Nukumaruan
Bottomsets
Foresets
Topsets
neritic
marginal neritic
marginal oceanic
oceanic
Lack of benthic data for cluster analysis, probably no shallower than upper bathyal
BRACKISH
Inner
Mid
Outer
SHELF
Upper
Mid
Lower
BATHYAL
outer shelf
Association UCs
outer shelf-uppermost bathyal
Association UN
upper bathyal
Association UQ
mid bathyal and deeper
Association US
upper-mid bathyal
Association SU
mid bathyal
Association UB
upper-mid bathyal
Association UE
mid bathyal
Association UCn
upper bathyal
Association UE
outer shlef-upper bathyal
Association C
upper-mid bathyal
Association US
upper-mid bathyal
Association UBBq
upper-mid bathyal
Association UBq
upper-mid bathyal
Association EU
mid bathyal
Association UC

Association UCs (sample interval – 500 m)
Co-dominated by *Uvigerina* (W4), and *Cibicides subhaidingerii* (W13; >20% for both), with a reasonably large contribution from *Lenticulina* sp2. (W13; 10-19.9%) with the rare specimen of *Karreriella cylindrica* (W2; 500 m). Very little diversity apart from these three main species. Outermost shelf to upper bathyal environment indicated.

Hayward (1984) noted few foraminifera between 165 and 500 m, though inner to mid shelf depths were inferred. Abundance of shallow water faunas in some samples (particularly in upper few hundred metres) suggests considerable reworking and re-deposition of inner shelf sediments to mid to outer shelf depths. The intervals in which shallow water taxa are found may be attributable to periods of lower sea level. Hayward (1984) is sceptical of the site ever being as shallow as inner shelf, and this study supports that idea (Fig. 6.14). Lack of benthic faunas mean that most samples between 1980-2250 m (with the exception of 2240 m) are not included in cluster analysis. The benthics that are present are consistent with bathyal depths (e.g., *Sigmoilopsis schlumbergeri*, *Melonis* sp., *Eggerella bradyi*), with water depths progressively decreasing uphole. Fewer imported shallow-water taxa are observed with increasing depth downhole.

(e) Composite cluster analysis (Fig 6.15)
As well as undertaking cluster analysis on individual wells the same programme was run on combined samples (with more than 40 specimens) from all wells, and on species with 50 or more proportional specimens. The results are displayed in the same fashion as for individual wells, and exhibit excellent grouping of both species associations, and faunal sample associations. Clear trends can be observed between the dominant species associations and their predicted environmental ranges, and faunal sample association groupings, reflecting the relative position of each well site from both the modern day shoreline, and the paleo-shoreline (refer section 6.6). Scrutiny of both dendrograms has resulted in the division of thirteen species associations, and twenty-six distinct faunal sample associations (aw = Arawa-1, ar = Ariki-1, k = Kora-1 and w = Wainui-1).

Species associations:
C1 – *Ammonia parkinsoniana* f. *aoteana* (brackish and very slightly brackish environments), *Elphidium charlottense* (inner to mid shelf, most abundant <20 m), *Haynesina depressula* (inner to mid shelf, c.0-75 m), *Nonionellina flemingi* (mid shelf-upper bathyal, and deeper, quieter

inlets), *Zeaflorilus parri* (shallower than 25 m), and *Quinqueloculina seminula* (inner shelf). Brackish to inner shelf (less than 50 m, and possibly less than 30 m).

C2 – *Astrononion novozealandicum* (inner and mid shelf, more abundant at outer shelf and bathyal depths), *Cassidulina carinata* (most abundant from 100-1500 m), *Evolvocassidulina orientalis* (inner and mid shelf, most abundant at upper bathyal depths), and *Notorotalia* sp1. (most abundant at inner shelf, and outer shelf-uppermost bathyal, depths). Mid to outer shelf depths (50-200 m).

C3 – *Notorotalia finlayi* (inner and mid shelf, deeper parts of enclosed inlets). Inner and mid shelf depths interpreted (<100 m).

C4 – *Astrononion kickinskii* (probable brackish to inner shelf depths), and *Cibicides* sp4. (inner shelf to abyssal). Wide range indicated, but based on seismic position of sample depths, probably outer shelf to mid bathyal.

C5 – *Bolivinita pliozea* (extinct; most abundant upper bathyal), *Bulimina striata* (common at bathyal depths), *Siphouvigerina* (most abundant mid and lower bathyal), *Oridorsalis tenera* (mid shelf to bathyal), and *Sphaeroidina bulloides* (mid shelf-abyssal, most abundant at outer shelf to upper bathyal depths). Upper bathyal (200-600 m).

C6 – *Karreriella cylindrica* only. Upper depth range calibrated (at different sites) to between 200 and 500 m. Given that this is the only species in this association, the range can only be given as no shallower than uppermost bathyal (>c.200 m).

C7 – *Cibicides deliquatus* (outer shelf-mid bathyal), *Hoeglundina elegans* (widespread at mid shelf to bathyal depths), and *Haeuslerella parri* (extinct; inferred upper to mid bathyal). Depths interpreted as upper bathyal (200-600 m).

C8 – *Notorotalia taranakia* (mid shelf to upper bathyal). Lack of other fauna in association results in a depth interpretation of mid shelf to upper bathyal (50-600 m).

C9 – *Bolivinita quadrilatera* (deep outer shelf-abyssal, most abundant uppeı-mid bathyal), *Uvigerina* (most abundant upper to mid bathyal), *Dentalina* spp. (most common upper to mid bathyal) and *Bulimina marginata* f. *aculeata* (upper to lower bathyal). Upper to mid bathyal depths (200-1000 m) indicated.

C10 – *Cibicides neoperforatus* (upper to lower bathyal), *Oridorsalis umbonatus* (most abundant at upper-lower bathyal depths), and *Gyroidina* sp1. (most abundant outer shelf to upper bathyal). Upper bathyal (200-600 m).

C11 – Inclusion of only one genera (*Cibicides* sp4.) results in a wide-ranging environmental interpretation, from inner shelf to lower bathyal (<50 –2000 m).

C12 – Presence of *Cibicides subhaidingerii* (outer shelf-mid upper bathyal) gives an outer shelf to mid upper bathyal depth range (100-400 m).

C13 - *Notorotalia zelandica* is found at inner to mid shelf depths, but more especially so at inner shelf depths, and as such suggests an inner shelf environment (<50 m).

Faunal sample associations:

Association BA (sample intervals – k1300, k1280, k1260, k1320, ar1880 m, ar2000, ar2040, ar2100)
Dominated by large populations of *Bulimina marginata* f. *aculeata* (C9), with good populations of *Astrononion novozealandicum* (C2). Ariki-1 samples also contain sizable contributions of *Notorotalia* sp1. (C2). Mixed upper-mid bathyal and inner-mid shelf signal. Depositional environment is interpreted to be upper to mid bathyal, with mixing of shallower water fauna.

Association AE (sample intervals – k1340, k1360, k1380, and k1420 m; = association AB2, Kora-1)
Dominated by fauna of species association C2, *particularly Astrononion novozealandicum*, *Evolvocassidulina orientalis*, and *Notorotalia* sp1. Upper bathyal depths.

Association E (sample interval – aw1400 m)
As with the previous association, but with a distinct lack of *Astrononion novozealandicum*. Good populations of *Evolvocassidulina orientalis*, as well as *Notorotalia* sp1, costate *Uvigerina* (C9) and *Cibicides* sp4. (C11). Outer shelf to upper bathyal depths interpreted.

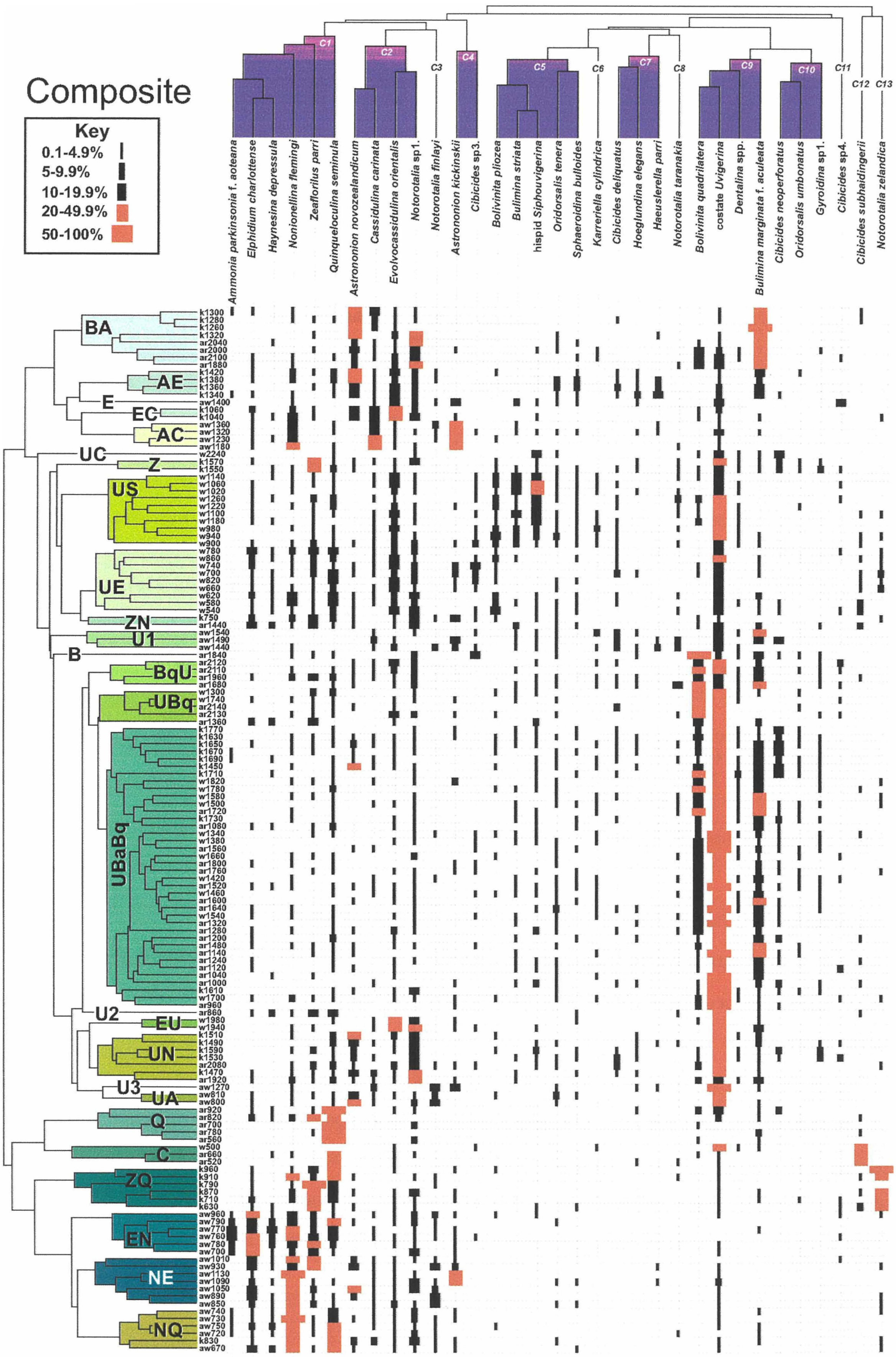


Fig. 6.15: Cluster analysis of benthic species, all wells. Expanded text on following page.

Environmental range for species associations:

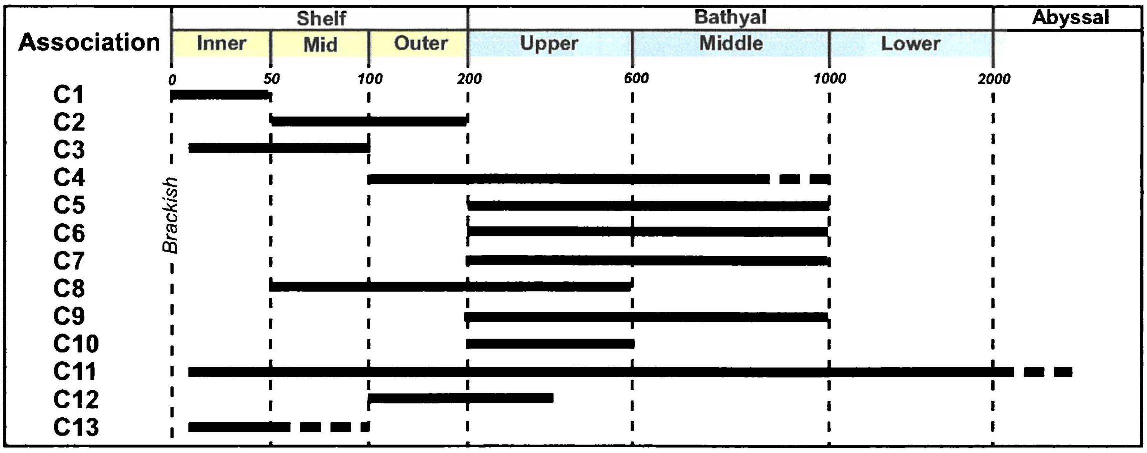


Fig. 6.15 continued: Composite cluster analysis on all samples with species that have more than 40 specimens and samples with more 50 specimens.

Association EC (sample intervals – k1060 and k1040 m; = association E, Kora-1)

Samples dominated by *Evolvocassidulina orientalis* (between 20 and 49.9%), with good populations of *Cassidulina carinata*, and *Astrononion novozealandicum* (all from species association C2). Quiet environments (outer shelf to upper bathyal) indicated.

Association AC (sample intervals – aw1180, aw1230, aw1320, aw1360and m; = association AC, Arawa-1)

Dominated by large populations of *Astrononion kickinskii* (C4), with large contributions of *Cassidulina carinata* (C2), and *Nonionellina flemingi* (C1). Outer shelf to upper bathyal.

Association UC (sample interval – w2240 m; = association Ucn, Wainui-1)

Sample is dominated by nothing specific, but does have reasonable populations of *Uvigerina* (C9), *Cibicides neoperforatus* (C10), and *Siphouvigerina* (C5). Mid bathyal depths or greater interpreted.

Association Z (sample intervals – k1550 and k1570 m; = association ZU, Kora-1)

Large populations of *Zeaflorilus parri* (C1; 20-49.9%), with reasonable numbers of costate *Uvigerina* (C9). Most other fauna in this association are connected to deeper environments. Mid bathyal, with considerable reworking and transport of inner shelf fauna.

Association US (sample intervals – w900, w980, w940, w1020, w1060, w1100, w1140, w1180, w1220, and w1260 m)

Dominated by large populations of costate *Uvigerina* (C9; >10%, and generally greater than 20%), with good populations of hispid *Siphouvigerina* also (C5). *Evolvocassidulina orientalis* (C2), *Bolivinita pliozea*, and *Bulimina striata* (C5) also occur in good numbers. Presence of *Stilostomella* at 900, 980, 1180, and 1260 m indicates mid bathyal depths for these sample intervals, with other depths being possibly shallower. Upper to mid bathyal depths interpreted, with some reworking and transport of mid shelf fauna.

Association UE (sample intervals – w540, w580, w620, w660, w700, w740, w780, w820, and w860 m)

Again dominated by *Uvigerina* (C9), but also large contributions of C1 and C2 fauna, including *Evolvocassidulina orientalis* (persistent), *Notorotalia* sp1, *Quinqueloculina seminula*, *Elphidium charlottense*, and *Zeaflorilus parri*. All sample intervals indicate an upper bathyal depositional environment, with considerable transport and mixing of inner to mid shelf fauna.

Association ZN (sample intervals – k750 and ar1440 m)

No one species dominates this association, although reasonable populations of *Zeaflorilus parri* and *Notorotalia* sp1. occur (from species associations C1 and C2 respectively). Faunal association is composed primarily of small populations of a number of shallow water taxa from C1 (e.g., *Elphidium charlottense*, *Nonionellina flemingi*, *Quinqueloculina seminula*). However, presence of deeper water taxa (*Astrononion novozealandicum* and *Evolvocassidulina orientalis*) indicate upper bathyal depths, with considerable reworking and transport of inner shelf taxa.

Association U1 (sample intervals – aw1440, aw1490, and aw1540 m)

One of three associations that have no characteristics other than reasonable to large populations of costate *Uvigerina*. Upper to mid bathyal depths inferred.

Association Bq (sample interval – ar1840 m)

Greater than 50% *Bolivinita quadrilatera* (C9), suggesting upper to middle bathyal depths, with presence of *Stilostomella* indicating a paleo-depth greater than 600-700 m (upper depth limit calculated at Ariki-1).

Association BqU (sample intervals – ar1680, ar1960, ar2110, and ar2120 m)
Large populations of *Bolivinita quadrilatera* (C9; up to and greater than 50%) with varying proportions of costate *Uvigerina*. Good contributions of *Bulimina marginata* f. *aculeata* (>5%) and presence of *Stilostomella* in many samples suggest mid bathyal depths.

Association UBq (sample intervals – w1300, w1740, ar1360, ar2130, and ar2140 m)
Uvigerina dominate over *Bulimina quadrilatera*, but same depth as previous association is also given to this (upper to mid bathyal).

Association UBBq (sample intervals – k1450, k1610, k1630, k1650, k1670, k1690, k1710, k1730, k1770, w1340, w1380, w1420, w1460, w1500, w1540, w1580, w1660, w1700, w1780, w1820, ar960, ar1000, ar1040, ar1080, ar1120, ar1140, ar1200, ar1240, ar1320, ar1280, ar1480, ar1520, ar1560, ar1600, ar1640, ar1720, ar1760 and ar1800 m)
Costate *Uvigerina* are ubiquitous throughout this large association, occurring in numbers greater than 20%, and often over 50%. *Bolivinita quadrilatera* and *Bulimina marginata* f. *aculeata* appear in varying proportions, from less than 5% to between 20-49.9%, but never occur in numbers over 50%. The presence of *Stilostomella* in many samples indicates lowermost upper to mid bathyal depths.

Association U2 (sample interval - ar860 m; = association U, Ariki-1)
Large (20-49.9%) population of *Uvigerina* (C9), with small populations of a number of species, predominantly from faunal sample association C1. Suggestive of deep outer shelf to upper bathyal depths, with reworking and transport of inner shelf fauna.

Association EU (sample intervals – w1940 and w1980 m; = association UE, Wainui-1)
Evolvocassidulina orientalis (C2) co-dominates with *Uvigerina* (C9), with good populations of *Notorotalia* sp1. (C2). Depositional environment is interpreted as upper to mid bathyal, with transport and mixing of shallower water fauna.

Association UN (sample intervals – k1470, k1490, k1510, k1530, k1590, ar1920, and ar2080 m)
Dominated by *Uvigerina* (C9), with large populations of *Notorotalia* sp1, and varying amounts of *Astrononion novozealandicum* (C2), with occasional occurrences of deeper water species e.g., *Stilostomella*, *Haeuslerella parri*, and *Bulimina marginata* f. *aculeata*. Outer to mid bathyal depths interpreted.

Association U3 (sample interval – aw1270)

Very large proportion of sample interval is dominated by *Uvigerina* (>50%) with good populations of both *Notorotalia finlayi* (C3) and *Astrononion kickinskii* (C4). Outer shelf to mid upper bathyal depths, but with some input from shallower water fauna inferred.

Association UA (sample intervals – aw800 and aw810)

Uvigerina is again the dominant species, but in lower numbers than previously, with *Astrononion novozealandicum* (C2) contributing between 10 and 50% of taxa. Outer shelf to upper bathyal depths.

Association Q (sample intervals – ar700, ar780, ar560, ar820, and ar920 m; = association Q, Ariki-1)

Apart from one sample (ar820), *Quinqueloculina seminula* (C1) contributes greater than 50% of the faunal sample association (and between 20-49.9% of ar820). Sample ar820 has high populations of *Zeaflorilus parri* (20-40.9%) and smaller contributions from *Elphidium charlottense* and *Nonionellina flemingi* (C1). Presence of deeper water faunas at 920 m (e.g., *Hoeglundina elegans*, *Cibicides neoperforatus, Cibicides wuellerstorfi*), and *Cibicides subhaidingerii* at 820 and 780 m, suggest deeper outer shelf to upper bathyal depths.

Association C (sample intervals – w500, ar520, and ar660 m)

Dominated by *Cibicides subhaidingerii* (C12), and *Quinqueloculina seminula* (C1; Ariki-1 samples only), or *Uvigerina* (w500). Outer shelf to upper bathyal (>c. 500 m based on presence of *C. wuellerstorfi* at ar660 m) depths indicated.

Association ZQ (sample intervals– k630, k710, k790, k870, k910, and k960 m)

Strongly dominated by *Zeaflorilus parri* and *Quinqueloculina seminula* (C1), with varying amounts (>50% for k960) of *Notorotalia zealandica* (C13). Mid shelf to upper bathyal depths indicated.

Association EN (sample intervals – aw700, aw760, aw770, aw780, aw790, and aw960 m)

Generally dominated by *Elphidium charlottense*, with strong populations of *Nonionellina flemingi*. Considerable contributions also from *Ammonia parkinsoniana* f. *aoteana* and *Zeaflorilus parri*, with smaller numbers of *Haynesina depressula* and *Quinqueloculina seminula*. All major contributors are from species association C1, with only sporadic occurrences of anything deeper than C2, and indicate mid to outer shelf paleodepths, shallowing to inner shelf.

Association NE (sample intervals – aw1010, aw930, aw1130, aw1090, aw1050, aw890, and aw850 m)
Nonionellina flemingi contribute the largest proportion to this association, with varying amounts of *Elphidium charlottense* (both C1). Varying and small amounts of other species from faunal association C1 (e.g., *Haynesina depressula*, *Quinqueloculina seminula*), and C2 (e.g., *Notorotalia* sp1, *Astrononion novozealandicum*) suggest a mid shelf to upper bathyal depositional environment, with transport of inner shelf faunas.

Association NQ (sample intervals – aw670, aw720, aw730, aw740, aw750, and k830 m)
Nonionellina flemingi and *Quinqueloculina seminula* (C1) strongly dominate this assemblage, with lower numbers of *Elphidium charlottense*, *Haynesina depressula*, *Zeaflorilus parri* (C1), and *Notorotalia* sp1. (C2). Small populations of *B. marginata* f. *aculeata* and *Cibicides subhaidingerii* in some samples indicate slightly deeper water depths. Uppermost bathyal for sample k830, otherwise mid to outer shelf depths, possibly shallowing to inner shelf depths.

From comparison with each of the individual wells, inferred paleodepths gained from the composite dendrogram correlate well with previously interpreted water depths. Associations delineated by combining all wells clearly show that, not only is there an overall decrease in paleodepth uphole at each site, but paleodepths also change in a lateral sense (i.e., increasing bathymetry with increasing distance from the modern day/paleoshoreline). Composite faunal associations have been included for comparative value on the paleobathymetry figures for each well.

(f) Mangaa-1

Mangaa-1, selected for use as a control well because of the recent in-depth study by Waghorn et al. (1996), provides a means of assessing in detail the paleobathymetry of the central axis of the Northern Graben. The resultant paleobathymetry is summarised in Fig. 6.16. Benthic faunal composition at this well is similar to or the same as observed at other sites. Hayward (1985b), observed that the interval from 450-1750 m is dominated by varying combinations of *Zeaflorilus parri* (457-981 m), *Elphidium charlottense* (457-1069 m), *Uvigerina rodleyi* (457-1750 m), *Nonionellina flemingi* (457-1216 m), *Notorotalia finlayi* (503-1069 m), *Astrononion novozealandicum* (960-1372 m), *Evolvocassidulina orientalis* (below 960 m), *Bolivinita quadrilatera* (below 1527 m), and *Bulimina aculeata* f. *marginata* (below 1600 m).

Fluctuating water depths between inner and mid shelf (450-960 m) and mid to outer shelf (960-1527 m) are suggested by both Waghorn et al. (1996) and Hayward (1985b), with an overall shallowing up-hole (bathyal-restricted taxa such as *Notorotalia profunda* and *Melonis barleeanum* are noted; Hayward, 1985b).

Benthic faunas are scarce in Tongaporutuan to Kapitean sediment, thus evidence is poor for assessing paleodepths. Sediment is inferred to have accumulated at bathyal depths. Opoitian paleodepths are more certain, with a change up-hole from lower bathyal to mid bathyal depths in the latter part of the Opoitian (faunal assemblages are dominated by *Uvigerina perigrina*, *Sigmoilopsis schlumbergeri*, *Uvigerina miozea*, *Gyroidina pseudorbicularis*, and *Cibicides neoperforatus*). Paleodepths progressively shallow through the Waipipian and Mangapanian, with assemblages dominated by *Bolivinita quadrilatera*, *Bulimina marginata* f. *aculeata*, *Oridorsalis umbonatus*, *Uvigerina perigrina* below 1975 m, while an upper bathyal depth is interpreted between 1907-1864 m (Waghorn et al., 1996). Hayward (1985b) notes that the fauna in a sidewall core at 1937 and 1788 m are more characteristic of outer shelf depths, and this may indicate a sea level low (or lows), which would have enabled sediment from shallower depths to be reworked or transported en-masse downslope. The interval from 1710-1829 m (lower Nukumaruan) is suggestive of deeper conditions (mid bathyal) than immediately down-hole, indicated by the presence of *Sigmoilopsis schlumbergeri* (Waghorn et al., 1996). Benthic assemblages through the remainder of the Nukumaruan indicate shallowing and considerable seaward transport of shallow-water taxa.

6.3.2 (v) Benthic species diversity

Species diversity of benthic foraminifera is a further method of assessing the depth range of sampled intervals. Diversity refers to the number of different taxa in an assemblage, and if one species is dominant, it is said to be abundant (Murray, 1991a). Simple diversity (the number of species in a sample, *S*) is sensitive to the addition of rarer species, implying that, in larger samples, we should intuitively expect more species (Murray, 1991a; Ottens and Nederbragt, 1992). Simple diversity measures therefore offer little resolution in terms of (paleo) environmental trends. Because of this, diversity needs to be measured using a statistical method that is not dependent on sample size. The two methods which are commonly used for foraminiferal data are the α index of Fisher et al. (1943), and the information function (Shannon-Wiener index), *H'* (e.g., Buzas and Gibson, 1969; Gibson and Buzas, 1973). The

Fisher index and Shannon-Wiener index are complimentary; α eliminates the effects of sample size, while H' gives an indication of the heterogeneity within a sample (Murray, 1991a). Another method, used to measure the distribution (or evenness) of the number of individuals per species, is the equitability measure, E (Buzas and Gibson, 1969). In general, while species diversity varies from place to place and from depth to depth in different regions (Gupta and Srinivasan, 1992) lower diversity samples are more characteristic of brackish or paralic environments, while diversity increases with increasing water depth, to a maximum diversity at about upper bathyal depths (Hayward et al., 1999).

Appendix 3d (part a) includes a summary of these statistics for all wells. Following the procedure used for faunal assemblages, all samples having less than 40 benthic specimens are not included in these statistical assessments. As counts between samples varied, total specimens and dominant taxa were normalised against one gram of sediment.

Fisher Alpha Index ($\propto$)

The Fisher Index (α) generates a value based on the number of species in a given sample relative to the number of individuals in that sample (Hayward, 1986a). This value can often be used to distinguish between shallow water samples and deeper water samples. The value can be derived from the equation:

$$\text{alpha } (\alpha) = n_1/x$$

(where x is a constant <1 and $n_1 = N(1-x)$, N being the number of individuals; Murray, 1990)

but is more often than not read straight off a graph (Fig. 6.17). In this study, the graph used is taken from Murray (1973) after Fisher et al. (1943). Murray (1991a) states that, within living assemblages, communities (biotopes) associated with unstable, variable and marginal marine conditions (e.g., high energy shorelines, or areas of large sediment input) have lower α values, while those in stable, less variable, normal marine conditions (e.g., quieter, deeper water) tend to display higher α values. Low α values (less than 10) are characteristic of shallower environments, such as inner shelf or paralic. Maximum values (α = 20-30) are recorded around upper bathyal depths (Hayward, 1986a).

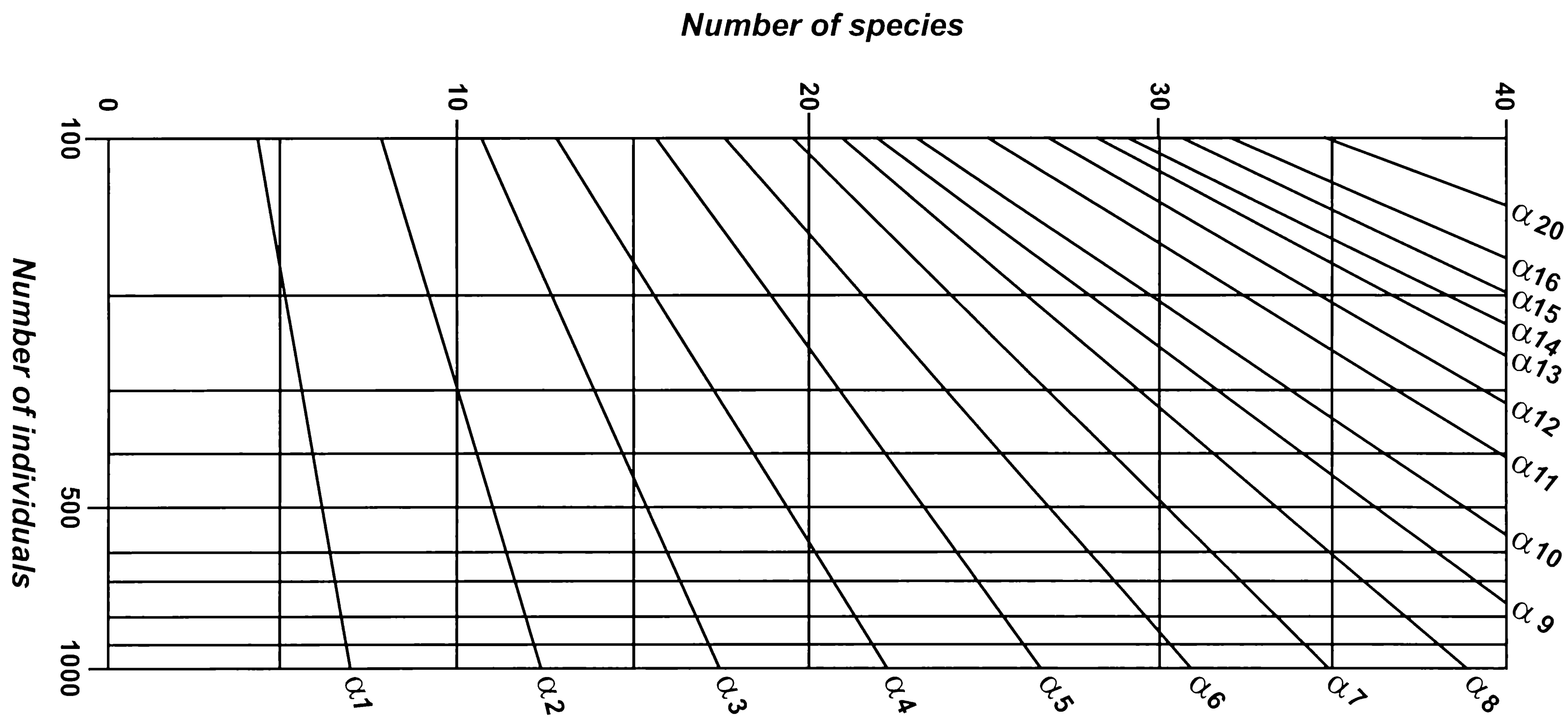


Fig. 6.18: Graph showing the relationship between species and individuals in an assemblage, and lines of equal diversity index (from Murray, 1973).

Values obtained for this diversity measure at Arawa-1 are all lower than 10 above 1400 m bKB (Fig. 6.18a), suggesting shallower, inner shelf environments above this depth, and deeper environments below this depth. This generally correlates well with the result obtained from faunal associations, which also indicate shallower (outer shelf or less) environments, and also with planktic percentage (<25%) and a corresponding neritic watermass. The higher α index observed between 1540 m and 1320 m coincides with lower total benthic numbers, and is consistent with deeper paleo-depths (outermost shelf-mid bathyal) and a slight increase in the planktic ratio. Throughout the Giant Foresets Formation at this site, benthic numbers dominate over planktic, but drop dramatically in the underlying Manganui Formation. Because of the very low benthic numbers recorded (less than 30), samples below 1540 m are not included in this (and other) diversity measures.

The Fisher index is more variable at Ariki-1 (Fig. 6.18a) and Kora-1 (Fig. 6.18b). Ariki-1 has α values ranging from less than 2 to 11, while Kora-1 displays α values ranging from less than 2 to greater than 12. While these values are lower than would be expected, particularly at Ariki-1, which displays more marginal-oceanic to oceanic water mass conditions, and deeper overall paleobathymetry, both wells show an overall shallowing trend up-hole from the base of the Giant Foresets Formation. Part of this disparity may be due to the difference in sieve size used (125 μm for Arawa-1, vs. 150 μm for the other three wells; see Schröder et al., 1987; Van der Zwaan et al., 1999). Smaller species that are able to slip through the 150 μm mesh, but not the 125 μm, which may increase the number of individuals, and therefore, the relative number of different species. Comparison of the number of specimens in a sample versus the number of species shows that there is indeed a direct correlation between these two variables. However, lower than expected values may also be attributed to a number of other factors, including taxonomic loss of specific species to dissolution or solution (though not really considered to be of importance at these sites), and environmental instability (e.g., at Ariki-1; Gibson and Buzas 1973).

Gupta and Srinivasan (1992) argue that environmentally unstable areas have lower evolutionary potential than stable environments, resulting in a decrease in species diversity. Seismic line P95-168 (see Appendix 6) clearly shows that the area in which Ariki-1 is located was subject to massive syn-sedimentary down-slope slumping throughout much of the latter part of the depositional history of the Giant Foresets Formation (degradational seismic facies). Sudden large influxes of sediments are catastrophic to benthic communities, resulting in rapid smothering of both infaunal and epifaunal taxa. An event such as mass wasting of the shelf margin understandably takes a certain period of time in which the substrate needs to stabilise, before new communities are able to re-colonise and establish themselves. Gupta and Srinivasan (1992) suggest that a young ecosystem will give rise to low diversity. If slumping or mass sediment transport was occurring on a regular basis (e.g., funnelling along a submarine canyon, slumping of the shelf-slope break), ecosystems may not have had time to establish, which would be reflected by lower than expected diversity values, such as seen at Ariki-1.

Figure 6.18b illustrates a distinctive and wide band of barren to near barren strata (from 1060-1240 m) within the Kora-1 sequence. Benthic numbers increase away from this zone. This zone approximately correlates to a distinctly channelised interval incorporating seismic units A37 and A38, and in the upper part of the barren zone, corresponds to the initial influx of coarser sediment to the site (Fig. 6.19). This gives confidence to the idea that specimens and species numbers (and therefore diversity indices) may be highly affected by on-going, and sometimes catastrophic, sedimentary processes. A lower barren zone, observed between 1770-1790 m bKB correlates to the base of the Giant Foresets Formation/top of the Miocene volcanics. This lower barren interval is lithologically distinct from the upper interval, and has been identified both in this study and that of Hayward and Strong (1993) to be totally devoid of any foraminiferal taxa.

Wainui-1 (Fig. 6.18b) is interesting because it displays the inverse situation to the three previous wells discussed – i.e., α values *increase* uphole – despite the fact that benthic specimens contribute 60-75% of the total population in most samples. Wainui-1 well site has always been in a more distal position relative to the nearest landmass, and under deeper water depths than most other sites during the Plio-Pleistocene. Greatest species diversity is usually recorded at upper bathyal depths (Hayward, 1986), and diversity of benthic foraminifera is related to (among other things) availability of food, watermass and bottom water conditions, and oxygen levels (refer to discussion in section 6.3.2(i)).

Fig. 6.18 (following pages): Compilation of diversity measures for Arawa-1, Ariki-1 (Fig. 6.18a), Kora-1 and Wainui-1 (Fig. 6,18b). Intervals in where there are few benthic specimens (less than 40) are not used in statistical analysis. Zones within the well section that are barren or nearly barren of benthic fauna are indicated.

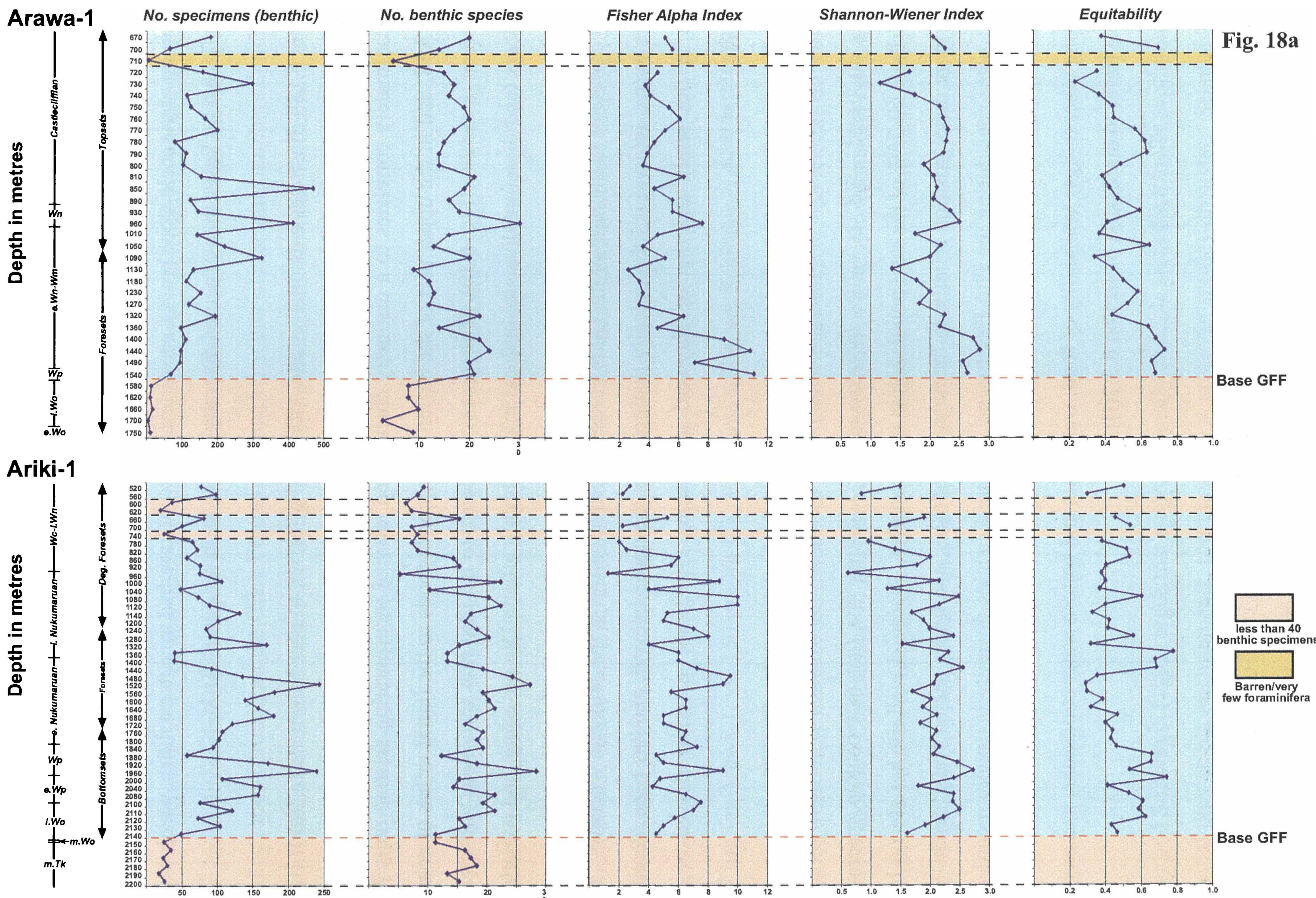
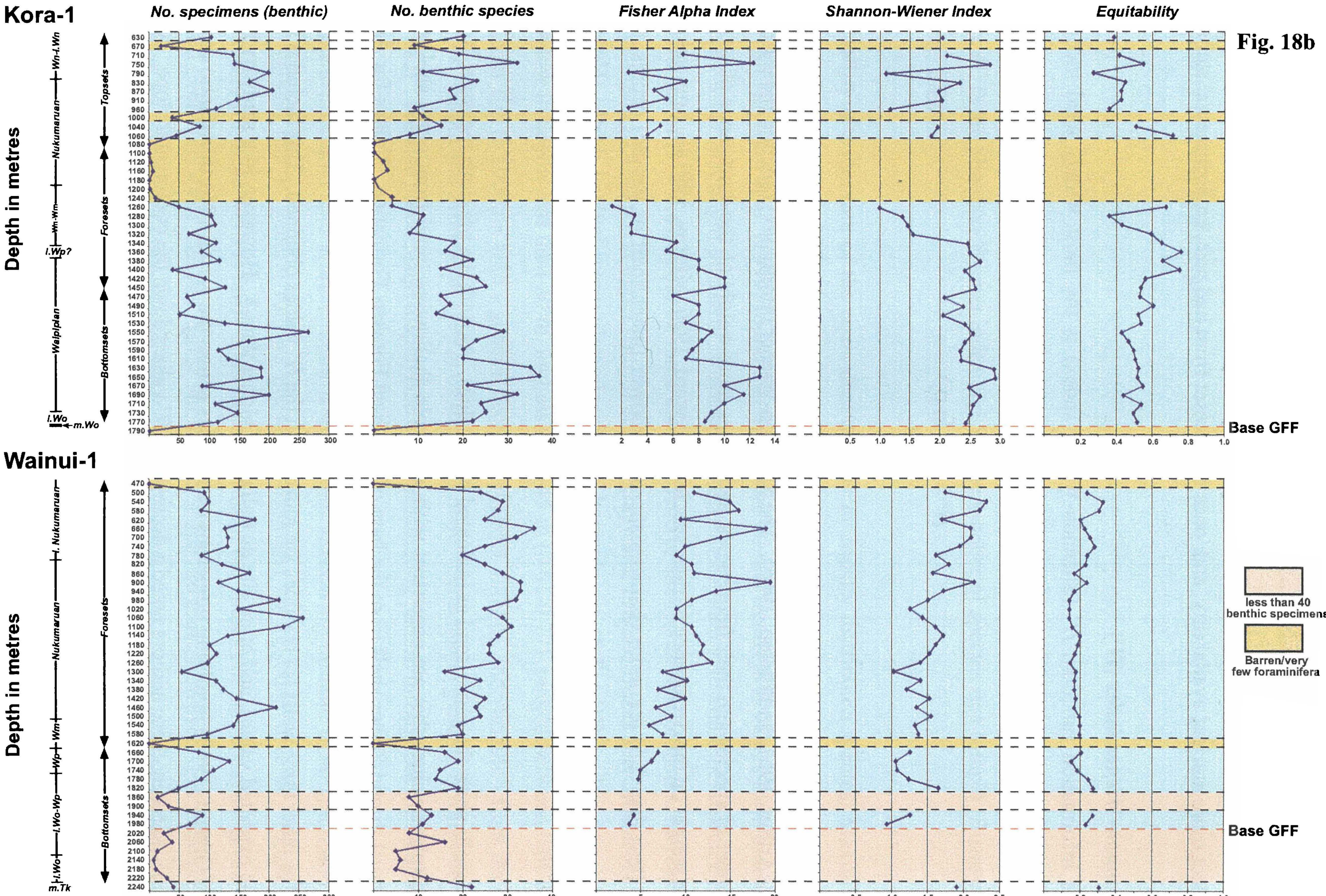

Arawa-1
Ariki-1
Depth in metres
No. specimens (benthic)
No. benthic species
Fisher Alpha Index
Shannon-Wiener Index
Equitability
Castlecliffian
Wn
e.Wn-Wm
Wp
l.Wo
e.Wo
Topsets
Foresets
Wc-l.Wn
l. Nukumaruan
e. Nukumaruan
m.Wo
m.Tk
Deg. Foresets
Bottomsets
Base GFF
less than 40 benthic specimens
Barren/very few foraminifera

Fig. 18a

Fig. 18b


Kora-1
Depth in metres
No. specimens (benthic)
No. benthic species
Fisher Alpha Index
Shannon-Wiener Index
Equitability
Topsets
Foresets
Bottomsets
Wn-l.Wn
Nukumaruan
Wn-Wm
l.Wp?
Waipipian
l.Wo
m.Wo
Base GFF
Wainui-1
l. Nukumaruan
Nukumaruan
Wm
Wp
l.Wo-Wp
l.Wo
m.Tk
less than 40 benthic specimens
Barren/very few foraminifera

The low diversity observed in Wainui-1 may be a result of the general paucity of sediment that reached this site during the Late Opoitian to Early Nukumaruan, low nutrient availability, and variably low surface productivity as evidenced by the planktic curve (Fig. 6.2). Values towards the top of the sampled sequence are more representative of bathyal conditions, and possibly reflect more nutrient availability, and, once the foreset front had past, environmental stability.

Shannon-Wiener index (H')

The information function, *H'* (Shannon-Wiener index) is the most widely used index for diversity measures used in biological research (Hayward, 1986a). It is defined as:

$$H' = \sum_{i=1}^{s}(\rho_i \ln \rho_i)$$

where ρ_i is the proportion of the *i*th species (from Hayward, 1986a, after MacArthur and MacArthur, 1961, and Pielou, 1966). The Shannon-Wiener information function measures the number of species and their frequencies, and places more weight on the dominance of taxa and less weight on the occurrence of rarer species (Hayward, 1986; Murray, 1991a). Low values of *H'* (less than 2.5) are more characteristic of shallower (inner shelf or paralic) depths, while maximum values (*H'* = 3-4) are obtained at approximately upper bathyal depths (Hayward, 1986a).

All wells show the same trend as indicated by α values. As with the previous measure, Arawa-1 displays values that are consistent with a up-hole shallowing trend from outer shelf-upper bathyal to mid-?inner shelf from the base of the Giant Foresets Formation. Values for Arawa-1 range from between 2.5 and 3 (1400 m to 1540 m) to less than 1.5 (1130 m). Trends observed in the other three wells for α values are mirrored in the *H'* values obtained. This includes lower than expected values, and at no site do values go higher than 3. However, in a study by Buzas and Gibson (1969) involving numerous transverses across the western North Atlantic shelf- to slope- to basin, these authors noted a decrease in *H'* values down the continental slope. After a peak in diversity at 100-200 m, where they recorded *H'* values of about 3.0, values slowly decreased down the slope to less than 2.0 at 2000 m, increasing after this to a maximum of 3.5 at 4977 m. Wainui-1 in particular seems to reflect this trend somewhat, with lower values recorded with depth uphole, and concomitant uphole decreasing paleodepths. Diversity values for Arawa-1, Ariki-1, and Kora-1 peak (between ~2.3 to nearly 3.0) at inferred outer shelf to upper bathyal paleodepths.

Equitability (E)

Equitability is given by the equation:

$$E = \frac{e^{H'}}{S}$$

Where H' is the Shannon-Wiener index and S is the number of species in a sample (Buzas and Gibson, 1969).

Equitability is a measure of the equivalent number of equally distributed species. Maximum values (E =1) are obtained in a completely even sample, where all species are equally frequent. Although Hayward et al. (1999) found that there were no real trends in E values from brackish to deeper water that might be useful in assessing paleoenvironments, it would be expected that lower values of E (< 0.4) would be more characteristic of shallower environments, increasing to a maximum (0.45-0.65) at upper bathyal depths (Hayward, 1986a). Buzas and Gibson (1969) found that E was more variable than either simple diversity (S) or H'. Their study observed that E was low in depths less than 30 m, highest at 35-45 m (0.61-0.77), ranged between 0.30 and 0.50 from 100 to 1000 m, with a general decrease at about 1000 m, and variable values deeper than this. Gupta and Srinivasan (1992) on the other hand, found that in general, E followed the trend of H', but also noted that it displayed a highly fluctuating pattern throughout the Neogene.

At Arawa-1, Ariki-1 and Kora-1, there is an overall shallowing up-hole trend (increasingly lower values) that in general follows the trend illustrated by the other diversity measures. Equitability for these wells is fairly consistent in terms of the values one might expect given the inferred overlying water masses. Some anomalous 'blips' do exist, however. For example, E increases up towards, and decreases away from, the wide barren zone seen in Kora-1. The increase in E is coincident with low specimen numbers on either side of this zone, and may be a reflection of the sedimentary processes that created this zone. Catastrophic sedimentary events (e.g., mass movement, turbidity current deposits) have the ability to completely smother and wipeout benthic communities. When such events occur, the first species to re-colonise this essentially 'new' ecological niche are those that are able to quickly capitalise on the opportunities present, and rapidly re-establish themselves (i.e., opportunistic species). Hence, equitability is higher because assemblages are dominated by large numbers of one or two species (refer 6.2.3(v)). This may also account for the high values of *E* observed in Ariki-1 between 1360-1400 m, which correlates to a channelised deposit interpreted from both wireline (Sf4) and seismic facies.

Wainui-1 displays values that are quite different to the other wells discussed. At this site, the smallest values of any well section are recorded, with a narrow band of values ranging from 0.15 to 0.3. Given that values approaching 1 indicate species are more evenly distributed, this suggests that representative specimen numbers vary widely over a range of different species. Comparison with the other two diversity indices illustrates that equitability values at Wainui-1 are highest where benthic specimen numbers are lowest. In a study of northern Indian Ocean fossil benthic faunas, Gupta and Srinivasan (1992) found that *E* showed a highly fluctuating pattern throughout the Neogene, and they related periods of high values to periods of more stable environments, and attributed low values to intervals of environmental instability. They suggested this instability was the product of intensification of deep-water circulation as a result of global climatic cooling, and widespread Antarctic glaciation. Scott and Crundwell (1994), suggest that where *E* is the inverse of the number of benthic specimens, this may indicate an increase in productivity that is connected to environmental stress. Under both these situations, one particular species may dominate over all others, and grow into a large population. This supports the idea that this site was subject to more variable conditions throughout the Plio-Pleistocene than any of the four well sections.

6.3.2 (vi) Species domination – a sedimentary or oceanic response?

The predominance of certain species, and the factors that control this, can aid in the interpretation of past depositional and environmental conditions. Particular species, such as *Uvigerina*, *Evolvocassidulina,* and *Cassidulina*, are what are termed 'opportunistic species' – i.e., they are able to colonise in conditions that would be adverse to the survival of many other species, such as low oxygen (stressed) environments (Gooday, 1988). The importance of these opportunistic species becomes apparent when trying to understand the controls that are acting on entire assemblages, and thus within the depositional basin. These controls may be physical (i.e., sedimentary) or physico-chemical (such as change in water mass characteristics). Sample intervals with low diversity (few species) but high specimen counts may be indicative of changing environmental conditions as the result of changing oceanic circulation patterns, and restriction of available niches (e.g., the stressed environments of Scott and Crundwell, 1994), or a reflection of the aftermath of a catastrophic sedimentary event. Conversely, samples with very high diversity values, and mixed shallow-water and deeper-water faunas, are more likely to be as a result of post-mortem mixing. These characteristics aid in the construction of paleogeographic maps (Chapter 7), helping to identify periods of uplift, of decreased or increased invigoration (related to disappearance or re-emergence of coastal and oceanic currents) and times of lower of higher relative sea level, with an associated increase in mass-flow depositional mechanisms (e.g., slumping, fan building).

To illustrate species dominance, the most abundant taxa from species assemblages, plus taxa representative of very shallow, or very deep environments, were plotted against depth, simple diversity, total benthic specimens, and texture (percent sand) uphole for each well (Fig. 6.19 a and b). Textural curves are useful in understanding oceanic conditions and fluctuations in current velocities through time by the trends these curves display (Allison and Ledbetter, 1982), and hence give an indication of the variables that have influenced benthic and planktic foraminifera behaviour and distribution.

All well sections show a clear trend of increasing dominance of shallow water taxa with decreasing paleodepth uphole. This trend reflects the decreasing distance from the slope margin through time, and the increasing influence of sediment transport from inner shelf environments to outer shelf and bathyal environments.

(a) Arawa-1

At Arawa-1, benthic numbers are low in the Manganui Formation, and correlate to an interval of coarser sediment textures, and wireline motifs indicative of mass-flow (Sf1) and turbiditic (Sf2) deposition. Upper to mid bathyal conditions, under fully oceanic water masses, were present. Benthic numbers rise sharply near the base of the Giant Foresets Formation (Fig. 6.19a), and coincide with finer sediment textures, shallowing paleodepths uphole, and consistently high numbers of specimens, particularly *Quinqueloculina seminula*, with lesser numbers of *Astrononion novozealandicum*, *Cassidulina carinata*, *Evolvocassidulina orientalis, Nonionellina flemingi,* and *Uvigerina.*

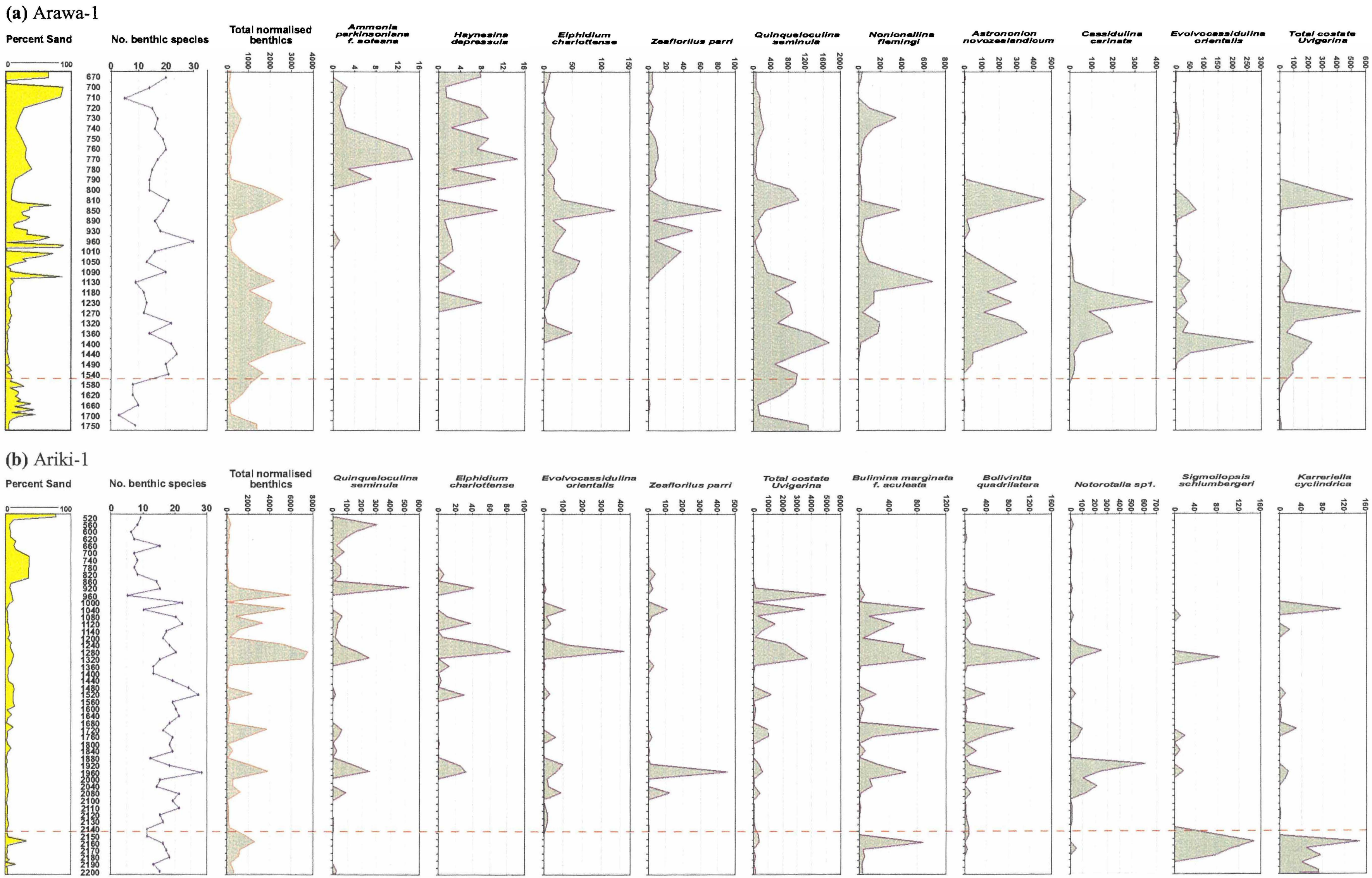
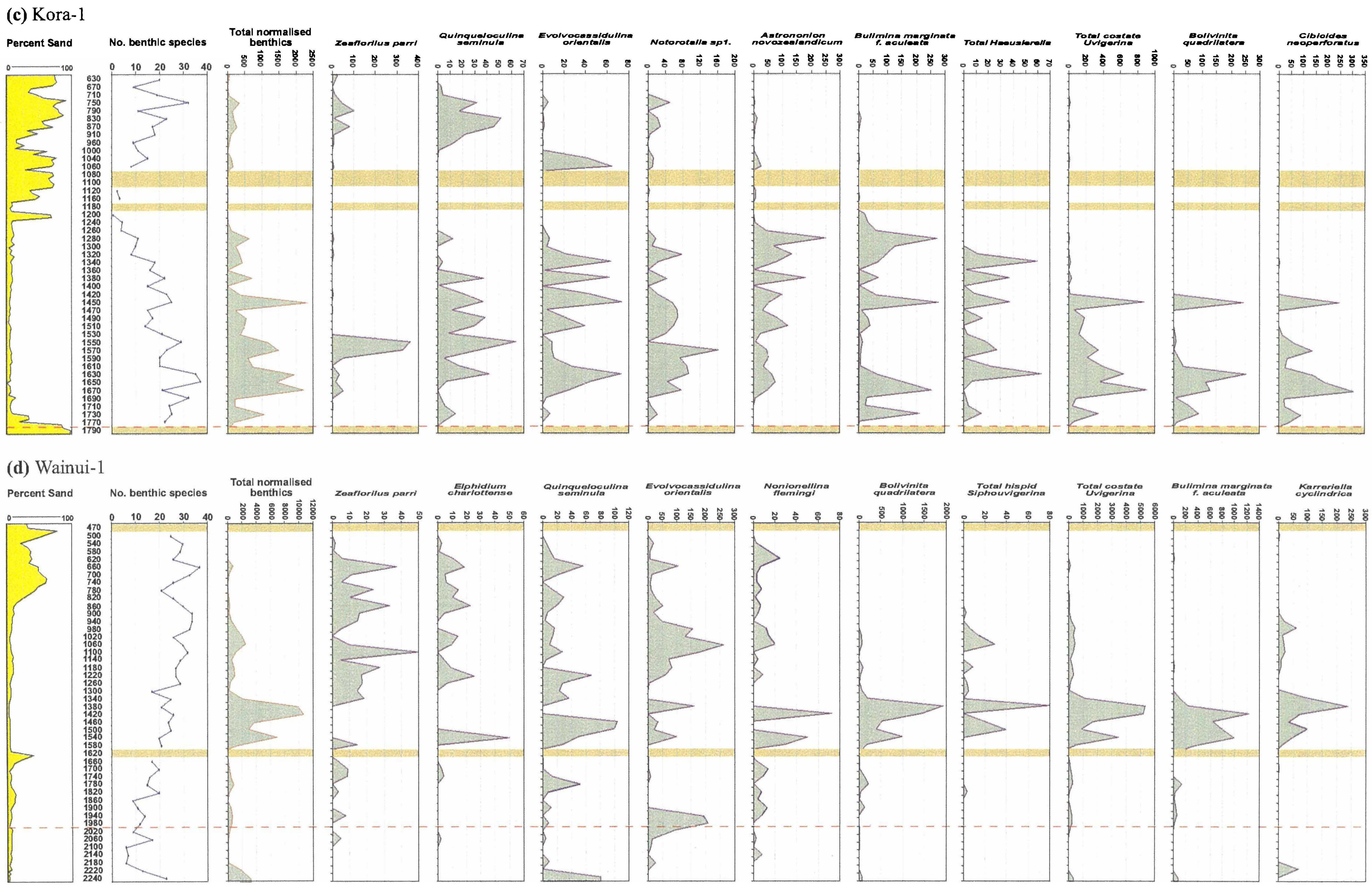


Fig 6.19 (a & b): Dominant species with depth (specimens per gram (normalised)). Zones which are totally barren of foraminifera are indicated by light brown ( ) colour. Base of the Giant Foresets Formation is delineated by dashed red line. Figure 19 (c & d) conitnued over page.


(c) Kora-1
Percent Sand
No. benthic species
Total normalised benthics
Zeaflorilus parri
Quinqueloculina seminula
Evolvocassidulina orientalis
Notorotalia sp1.
Astrononion novozealandicum
Bulimina marginata f. aculeata
Total Haeuslerella
Total costate Uvigerina
Bolivinita quadrilatera
Cibicides neoperforatus
(d) Wainui-1
Percent Sand
No. benthic species
Total normalised benthics
Zeaflorilus parri
Elphidium charlottense
Quinqueloculina seminula
Evolvocassidulina orientalis
Nonionellina flemingi
Bolivinita quadrilatera
Total hispid Siphouvigerina
Total costate Uvigerina
Bulimina marginata f. aculeata
Karreriella cyclindrica

Abundance spikes are observed at 1400 m (dominated by *Quniqueloculina seminula*, *Evolvocassidulina orientalis,* and *Uvigerina*), 1320 m (dominated by *Uvigerina*), 1270 m and 1230 m (dominated by *Cassidulina carinata*), and 1130 m (dominated by *Nonionellina flemingi* and *Astrononion novozealandicum*). These abundance spikes appear to be related to slightly coarser sediment textures, and may be a faunal response to a sedimentary event. Often the deeper water species such as *Cassidulina carinata*, *Evolvocassidulina orientalis*, and *Uvigerina* spp. contribute low background numbers only.

Shallow water taxa (e.g., *Haynesina depressula, Elphidium charlottense, Zeaflorilus parri,* and *Ammonia parkinsoniana* f. *aoteana*) are most abundant above 1060 m bKB, suggesting a greater component of transport and deposition of inner shelf faunas, via sediment flows, from the inner shelf to mid to outer shelf depths. Large abundance spikes dominated by *Uvigerina*, *Evolvocassidulina orientalis*, and *Cassidulina carinata* between 820-810 m (immediately above a textural spike), and the sporadic occurrence of *Uvigerina* up-hole (to ~ 700 m bKB), supports a minimum mid shelf depth limit.

Generally, peaks displayed by the shallow-water taxa coincide not only with each other, but also with higher sand percentages, and lower overall benthic numbers. This suggests that either (a) environments are very shallow, and occur within the shore-connected sediment wedge; or (b) faunas are transported to deeper areas as part of coarse-grained sediment flows, derived from inner shelf regions, possibly during periods of lower sea level. An abundance of shell hash is also recorded in samples above 1000 m. This supports the idea of an inner shelf origin, but suggests (due to lack of whole shells) some degree of transport. Low total benthic numbers and appearance of very shallow water species indicates possible dilution by a rapid influx of sediment, and reinforces the idea that some transport has been involved. Given that the inferred paleo-depths in the sampled section have not been shallower than at present, reworking and mobilisation of sediment derived from shallower environments, and deposited in deeper-water environments, is the preferred explanation.

(b) Ariki-1

Uvigerina often make up at least a few percent of the total assemblage at Ariki-1, and at times, up to 84%. Higher planktic totals, overall more oceanic water mass conditions, and dominance of deeper water taxa (Fig. 6.19b) reflect the deeper position of this site. Here, low simple diversity corresponds well with low total benthic numbers, suggesting that there was no real dominance by just a few species. However, a number of spikes are distinctly obvious. Peaks observed lower in the sequence (between 1320-920 m bKB) include large populations of *Evolvocassidulina orientalis*, *Uvigerina*, *Bulimina marginata* f. *aculeata*, and *Bolivinita quadrilatera*, together with lower numbers of *Quinqueloculina seminula* and *Elphidium charlottense*, and correspond to more muddy sediment textures. An abundance of shell hash was also noted within this interval. The predominance of deeper water fauna, together with the shallow water faunal component, implies that a degree of transport from inner shelf environments is still a factor in faunal assemblages. However, the very high numbers of *Uvigerina* (up to 5000/gram), often coupled with smaller peaks in *E. orientalis* numbers, suggest that the depositional environment at times may have been oxygen stressed. Wireline facies motifs (hemipelagic and numerous small-scale upward-fining Sf2 units), plus position (mid-lower slope) as indicated by seismic, support a site subject to frequent passage of turbidite flows. Peaks observed higher in the succession (above 860 m bKB) include the dominant shallower water taxon *Q. seminula*, and correspond to coarser and shelly lithologies, and the degradational seismic facies, indicating (a) considerable transport from inner to mid shelf regions; and (b) downslope reworking through mass failure mechanisms.

Seismic reflection profiles illustrate that sediment began piling into the vicinity of Ariki-1 and surrounding areas in the Waipipian, accelerating through the Nukumaruan and Castlecliffian. Initially, sediment was deposited with minimal disruption to underlying strata. However, as sedimentation rates accelerated, over-steepening and destabilisation of the slope occurred, resulting in erosion of the underlying sediments, and massive down-slope slumping. Continual dumping of sediment on the basin floor and slope would create variable conditions, often not conducive to establishing flourishing benthic communities. It is suggested here that continual sedimentation smothered deepwater communities, creating dysoxic or anoxic environments, while introducing shallow water fauna and a sudden increase in organic (nutrient) flux (e.g., Alve and Murray, 1997). The first species to recolonise were the infaunal *Uvigerina*, capitalising on the abundance of food, and lack of competition. This opportunistic benthic foraminifera was later joined by other bathyal taxa (such as *Bulimina marginata* f. *aculeata*, and *Bolivinita quadrilatera*, followed by *Notorotalia* spp.), re-establishing a respectable community, before once again being buried by the next influx of shelf-derived sediment. Benthic peaks correspond to, or closely follow, lower planktic ratios, and are possibly linked to lowering sea levels throughout the Plio-Pleistocene.

(c) Kora-1

A clear trend up-hole between dominant benthic species and decreasing paleo-depths is illustrated in Fig. 6.19c, with *Cibicides neoperforatus*, *Bolivinita quadrilatera*, *Uvigerina*, *Haeuslerella* spp, *Bulimina marginata* f. *aculeata* and *Astrononion novozealandicum* dominant below 1200 m bKB, and *Evolvocassidulina orientalis*, *Quinqueloculina seminula*, and *Zeaflorilus parri* common above 1100 m. *Uvigerina* is a persistent benthic species identified throughout the entire sampled Kora-1 sequence, but are only really dominant below 1420 m, comprising up to 55% of the benthic faunas. *Haeuslerella* spp., though relatively abundant in the lower part of the sequence (below 1340 m), is last noted at 1320 m. Its abrupt departure from faunal assemblages may be related to the extinction of this genus during the Nukumaruan, rather than changing bathymetric conditions. An abundance spike at 1550-1570 m, dominated by *Z. parri* and lesser amounts of *Q. seminula*, does not correspond to any textural spike, nor does seismic character of this zone indicate anything in particular such as a channelised zone. However, wireline facies motifs suggest it is related to turbidite deposition, clearly indicating transport from the inner to mid shelf zone.

Barren and nearly barren zones (1790-1780 m and 1250-1060 m,) correlate well with coarser sedimentary textures. Lack of faunas in the upper barren zone suggest that either sedimentation was occurring too rapidly for benthic communities to exist, that bottom conditions were changing too rapidly for evolution (of the benthos community) to keep pace, taphonomic dissolution, or that the substrate was not conducive to benthic survival (possibly anoxic?). An abundance of shell hash and the occurrence of small indurated pebbles within an otherwise muddy sequence, coupled with a possible decrease in paleo-depth through the barren interval (1250-1060 m), is indicative of a rapidly changing environment, with an associated sudden influx of sediment. *Evolvocassidulina orientalis* is the first species to re-inhabit the sediment, apparently making use of the initial lack of competition provided by other taxa. The lack of faunas in the lower (1790-1780 m bKB) barren zone probably reflects the basic nature and texture of the volcaniclastic substrate, mineralogically being initially too alkaline to support benthic communities. Between these two zones, dominant species reflect both an overall upward shallowing paleodepth, and possible cyclical sea level changes as indicated by shallow water abundance peaks.

(d) Wainui-1

As with the previous wells, shallower water species occur more frequently with decreasing depth uphole. Total benthic numbers are very low within the Ariki Formation and lower part of the Giant Foresets Formation, reflecting its more distal position, and possibly lowered nutrient availability and surface water productivity. Benthic species do not occur in any great number until ~1580 m, above a barren zone between 1620 and 1640 m. Two exceptions are *Evolvocassidulina orientalis*, which contributes 36 and 28% of the total benthic assemblage at 1980 and 1940 m respectively, and *Uvigerina* sp., which contributes 39 and 30%. The dominance of these two species, and lack of any real contribution from other species, implies that these species prospered where others could not. The textural curve suggests that sediment flux was not the cause of this bloom.

The barren zone between 1620-1640 m bKB is associated with a marked textural peak, and has been described in the Well Completion Report (Shell BP Todd Oil Services Ltd., 1982) as being characterised by shell hash and pebbles. It is delineated on wireline logs by lower gamma-ray and density values, and on seismic section occupies a lower slope position. The dominance of opportunistic species immediately above this zone (e.g., Uvigerina and *Bulimina marginata* f. *aculeata*), as well as several shallow water species (e.g., *Elphidium charlottense*, *Zeaflorilus parri*, and *Quinqueloculina seminula*), indicate ongoing (catastrophic?) transport from shelf to slope. Paleobathymetric trends have indicated that this site was never shallower than outer shelf depths at any stage through the Wanganui Series, so it is reasonable to assume that these inner shelf faunas were transported and redeposited at bathyal depths.

Sample intervals characterised by shell hash and pebbles are noted at 1100 m, 920 m, and between 810 and 600 m, and a second barren interval is observed at 470 m (highest sample interval analysed). The four lower intervals correlate to peaks associated with shallow-water taxa, and to low overall total benthic numbers. As with other sites, increase in the number of shallow water taxa and coarser textural trends indicate increasing communication between this site and shallower shelfal environments. At Wainui-1, this trend is observed higher in the sequence, consistent with its more distal locale relative to the contemporary land mass and advancing progradational wedge.

6.3.3 Summary – changing paleoenvironments through time

Tongaporutuan – Kapitean (11.3-5.2 Ma)

The northern Taranaki Basin remained at bathyal water depths throughout the Tongaporutuan and Kapitean, with the contemporary shoreline in the vicinity of Manutahi-1 and Parakino-1, Wanganui Basin, extending east towards the modern Ruahine Range (P. Kamp, pers. comm. 2002). The northern and western parts of the study area were subject to sediment starvation (Ariki Formation), supported by the absence of shallow water faunas and dominance of deeper water faunas such as *Karreriella cylindrica* and *Sigmoilopsis schlumbergeri* within Ariki Formation samples from Ariki-1.

The study area remained at bathyal depths, under fully oceanic watermass conditions throughout the Late Tongaporutuan to Kapitean, with middle to lower bathyal depths recorded at Ariki-1, Tangaroa-1, and Wainui-1, and upper to middle bathyal depths at Arawa-1. Mid to lower bathyal depths (shallowing uphole) were recorded at Mangaa-1, Kahawai-1, Te Kumi-1 (Waghorn et al., 1996), and Awatea-1 (Strong et al., 1996). High planktic foraminiferal percentages (often greater than 90%), and large specimens indicate marginal oceanic to fully oceanic watermass conditions at all these sites during the Late Tongaporutuan to Early/middle Kapitean. At many sites along the eastern margin of the study area (e.g., Turi-1, Pluto-1, Awakino-1, and Pukearuhe-1), most of the Late Tongaporutuan to Pleistocene sequence has been subsequently eroded, so paleobathymetry can only be inferred. However, Late Tongaporutuan Urenui Formation sediments are present at Uruti-1, indicative of deposition in a slope environment, while foraminiferal analysis of Early Kapitean sediment preserved at Okoki-1 indicates that this site had shallowed to outer shelf by this stage (Crowley and Crocker, 1989).

Opoitian (5.2-3.5 Ma)

During the Opoitian there was a continuation of terrigenous sediment starvation throughout much of the northern and western parts of the study area, resulting in a paraconformity being developed between middle/Late Kapitean strata and mid/Late Opoitian strata, while to the south, in the vicinity of Arawa-1 and Taimana-1, voluminous mudstone deposition was still occurring. All sites in the northern and southern regions of the study area were at bathyal depths, under fully oceanic watermass conditions. This changed during the Late Opoitian, evident in a dramatic drop in planktic percentage noted at many sites, and a corresponding change from oceanic to marginal neritic/neritic watermass conditions. Benthic faunal assemblages suggest a change in water depth over this period, although because of the presence of the paraconformity, this change may not be as marked as illustrated. In fact, the paleobathymetric curve for Arawa-1 suggests a much more restrained alteration in paleobathymetry, probably related to the advancing progradational wedge.

Paleodepths at Mangaa-1 and Awatea-1 were deeper than elsewhere in the study area, remaining at middle to lower bathyal depths throughout the Opoitian and into the Waipipian (Strong et al., 1996; Waghorn et al., 1996). This is interpreted to reflect early development of the Northern Graben, with subsidence initially outpacing sedimentation.

Waipipian (3.5-2.79 Ma)

Water depths shallowed progressively in northern Taranaki Basin during the Waipipian Stage, from lower bathyal to upper bathyal, with rapid and focused sedimentation during the Waipipian resulting in the continual aggradation of voluminous terrigenous sediment in the Northern Graben (Mangaa Formation). Common occurrences of shallow water faunas in this interval, e.g., *Zeaflorilus parri* and *Ammonia parkinsoniana* f. *aoteana* (Hayward, 1985b; Strong et al., 1996), demonstrate efficient transport of shallow water faunas across the shelf into deeper environments.

By the end of the Waipipian the foreset front had prograded past Arawa-1 and Taimana-1, resulting in a gradual shallowing in the southern part of the study area. At more distal sites to the north, water depths remained between middle and upper bathyal depths, although mixed outer shelf faunas with deeper water species at Ariki-1 and Wainui-1 demonstrate that redeposition of shelf-derived sediment continued to occur. Tangaroa-1, Te Kumi-1 and Kora-1 also shallowed to upper bathyal depths during this Stage, and Kahawai-1 had reached outer shelf depths by the end of the Waipipian (Waghorn et al., 1996). Planktic percentages at many sites fluctuate widely through the Waipipian and into the Nukumaruan, and may be due to increased nutrient availability in the surface waters, dilution due to increasing terrigenous sediment input, and/or post-mortem addition of transported (shelf-derived) taxa.

Mangapanian (2.79-2.28 Ma)

Sites at which Mangapanian strata have been identified illustrate a continuation of the shallowing trend observed through the Waipipian. Eastern and southern sites shallowed to shelf depths, with outermost shelf to mid shelf depths recorded at Kahawai-1 (Waghorn et al., 1996), with Kora-1 and Arawa-1 localities attaining upper bathyal to outermost shelf depths, under marginal neritic and neritic watermasses, respectively. Wainui-1, Tangaroa-1, and Te Kumi-1 sites remained at middle to upper bathyal depths, with slightly shallower depths recorded at both Mangaa-1 and Awatea-1. Mangapanian strata are not identified at Ariki-1. Waghorn et al. (1996) note a marked increase in benthic diversity and abundance at the top of this stage, which they suggest reflects a change in bottom water conditions. A similar trend is observed at Wainui-1, and similarly probably reflects increased invigoration as a result of coastal upwelling systems related to proximity of the foreset front.

Early Nukumaruan (2.28-~1.6 Ma)

The lower Nukumaruan was a period of intense and voluminous sedimentation in the Northern Graben, recording the progression of rapid progradation and aggradation of the Giant Foresets Formation across the more northern regions of Taranaki Basin. Sites more proximal to the contemporary landmass (e.g., Awatea-1, Kora-1, Kahawai-1, and Mangaa-1) continued to shallow rapidly as the progradational front moved past each site, while the more distal sites (Ariki-1, Wainui-1, Tangaroa-1, and Te Kumi-1) remained at bathyal depths. Increased mixing of inner shelf to brackish taxa (e.g., *Zeaflorilus parri*, *Elphidium charlottense*, and *Ammonia parkinsoniana* f. *aoteana*) with deeper outer shelf to bathyal species demonstrates improved connectivity between the shallow neritic environment and the deeper bathyal environment, with sediment probably transported along one of the many channel systems mapped across the area.

Late Nukumaruan to Castlecliffian (~1.6-0.33 Ma)

Late Nukumaruan to Castlecliffian samples at all sites include an increasing content of inner shelf species, coarse lithologies, and large amounts of shell hash. At all sites this suggests more efficient sediment pathways between the shore-connected wedge and the upper slope/outer shelf. As discussed in Chapter 5, most of the terrigenous sediment supplied to the prograding continental wedge occurred during periods of low sea level, when water depths may have shallowed by as much as 80-100 m. By the Late Nukumaruan, even the most distal sites had attained uppermost bathyal to outer shelf depths (c. 150-400 m water depth). Any significant shallowing due to glacio-eustatic mechanisms would have moved the shoreline basinward many tens of kilometres, facilitating the transport of inner shelf sediment to outer shelf and bathyal depths. Frequent fluctuations observed in the paleobathymetric curves generated by this study suggest that oscillations in global sea level began to have a more obvious effect across the study area as the region progressively shallowed through the Nukumaruan and Castlecliffian. These oscillations are also noted at Tangaroa-1, Te Kumi-1, Mangaa-1 (Waghorn et al., 1996) and Awatea-1 (Strong et al., 1996) and would probably be found in other well sections with further (detailed) analyses.

6.4 Reconstructing past environments

Throughout this chapter, various aspects of the ecology of benthic and planktic foraminifera have been discussed in relation to the methods utilised to understand the controls on their distribution in the study area. This section undertakes to synthesize these aspects to better understand how this information can be used to help document the Pliocene to Pleistocene progradation and aggradation of the Giant Foresets Formation in northern Taranaki Basin.

An important concept behind using foraminifera to reconstruct depositional environments is the recognition that foraminiferal assemblages change over geologic time scales. These occur in response to changes in bottom water properties, organic flux from surface waters (Gooday, 1994), or postmortem alteration through addition of environmentally different faunas or taphonomic changes by diagenesis and dissolution (Hayward, 1986a). The last factor (taphonomic changes by diagenesis and dissolution) is not considered to be significant in the study area. More significant, given the high sedimentation rates throughout the Plio-Pleistocene, is the impact of sedimentary disturbance or oceanic changes on benthic ecosystems, and what this means in terms of (particularly) sea level change, and sediment source areas. To this effect, a systematic approach needs to be taken in the identification of intervals that have assemblages that are the product of either changes in oceanicity (with more productive oceanic currents sweeping over a site, or as the result of substantial depth changes), or a result of catastrophic and environmentally destructive sedimentary events.

A quick review

So far, the information gained from this study has accentuated a number of elements:

1. planktic foraminifera are excellent indicators of overlying surface water mass conditions, and indicate a major change in water mass conditions occurred between the Kapitean and Opoitian Stages;

2. benthic foraminifera, when analysed using both qualitative and quantitative methodologies, can give reasonably good estimates of paleo-depth;
3. paleobathymetric curves for the five control wells indicate an overall upward-shallowing trend through time;
4. influxes of shallower-water faunas into deeper environments are, in general, associated with coarse textural peaks, and in some cases, inferred perturbations in the paleobathymetric curve;
5. opportunistic faunas appear to proliferate in the interval immediately above influxes of coarser sediment;
6. low diversity values indicate periods of environmental instability and high stress;
7. coarse textural peaks correlate with lower planktic ratios.

A model for identifying sea level change

A model, using foraminiferal information, for identifying periods of lower relative sea level during the Plio-Pleistocene has been constructed using a number of assumptions developed from the observations listed above. The basic premise for the sea level change model hinges on the idea that periods of high sedimentation (e.g., mass flows, turbidites) are related to sea level change (lowering). When sea level lowers and shorelines move seaward, the ability for coarser-grained lithologies, sourced from more energetic shallower water environments, to be transported to deeper, less turbid environments, is significantly increased. During these intervals, not only does the postmortem transportation of shallow water faunas escalate, but also the associated influx of organic carbon allows the proliferation of opportunistic species (e.g., Hayward et al., 2002). Often, in the very deep environments, coarsening trends are not clearly illustrated on the textural curve, but shallowing conditions may still be indicated by the inclusion of (significant numbers of) shallow water taxa in deeper water faunal assemblages. In some situations, continual sediment disturbance may not be enough to register on the textural curve, but might be indicated by the tailing-off of benthic numbers (e.g., through the barren/ nearly barren zone in Kora-1). After a substantial depositional event occurs, which smothers all faunas, followed by a quieter interval, populations begin to stabilise. As a general rule, less sand is equivalent to less disruption of the benthic community.

The second part of the model incorporates planktic percentages. High planktic percentages indicate periods of high surface productivity, generally related to the patterns of current circulation and upwelling in the deeper oceans. However, in the Tasman Sea, the modern oceanic fronts, and associated current circulation pattern (Fig. 6.20) are inferred to have been relatively stable since the Early Miocene (Nelson and Cooke, 2001; Cooke, 2002). At around 7 Ma, Southern Component Mode Water (200-700 m) and Southern Component Intermediate Water (700-~1300 m) were affecting DSDP Site 593, located on the Challenger Plateau to the south west of the study area (Fig. 6.20; Cooke, 2002).

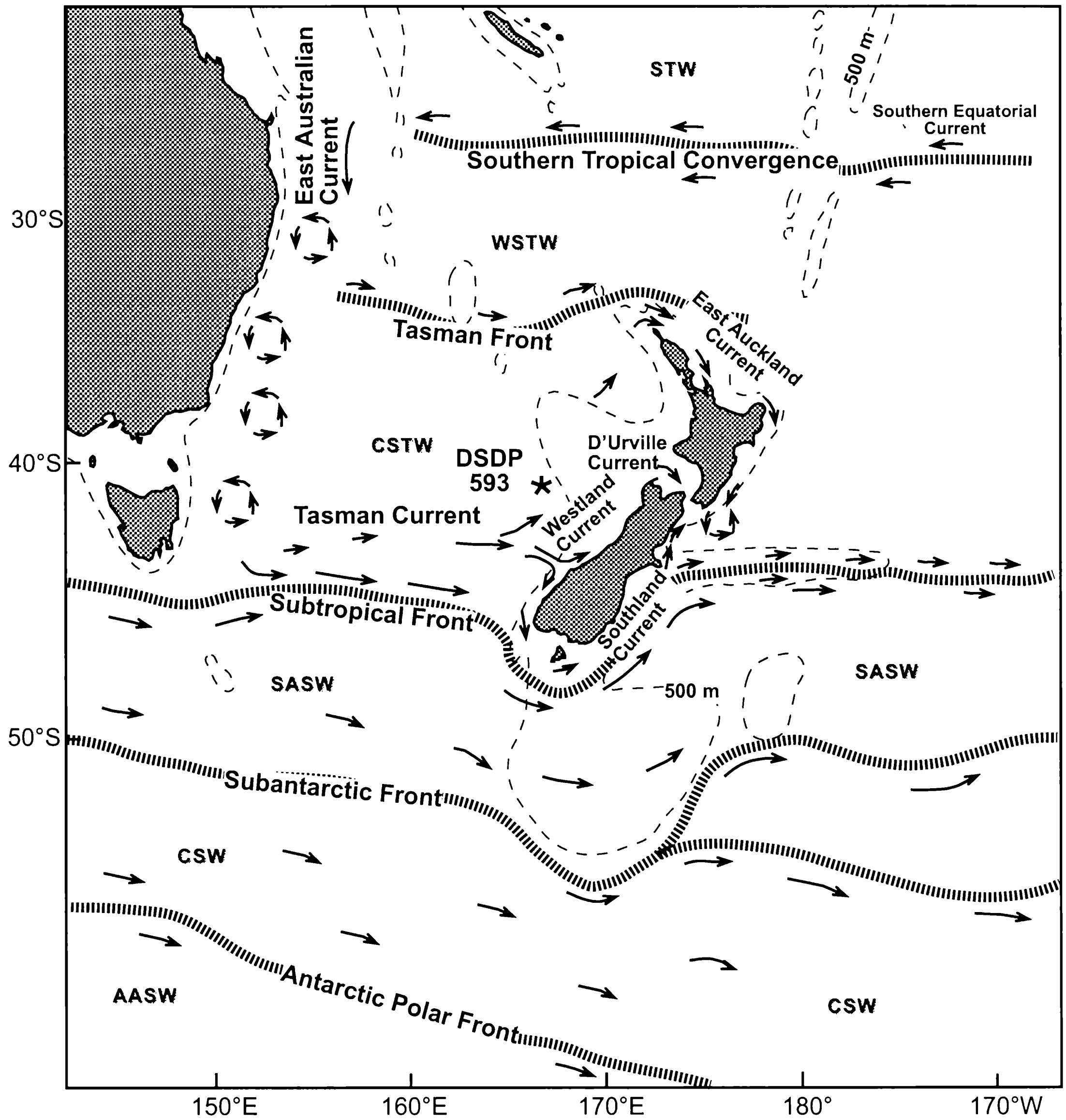


Fig. 6.20: Present day surface oceanographic features, and direction of deeper ocean currents. DSDP Site 593 is located immediately south of the Challenger Plateau (dashed line). Oceanographic features are not considered to have altered much since the Miocene (from Cooke, 2002). For abbreviations see Cooke (2002).

Surface current patterns were similar to modern day patterns, with the geostrophic Tasman Current (which extends to about 1100-1500 m water depth) pushing across the southern part of the Tasman Sea along the Subtropical Front (Stanton, 1973) towards the southern part of New Zealand, before splitting to form the southward flowing Southland Current, and northward-flowing extension of the wind-driven Westland Current. The Westland Current also splits, creating the D'Urville Current (flows through Cook Strait), and a more offshore current (still referred to as the Tasman Current) that continues northwards to rejoin the flow along the Tasman Front around northern New Zealand (Cooke, 2002). Importantly, there are no major upwelling zones in the modern Tasman Sea (Bradford and Roberts, 1978), nor have there been since the Miocene (Cooke, 2002).

The fluctuating planktic percentages and corresponding water mass conditions observed in all four well sections through the Plio-Pleistocene could thus be attributed to two primary causes; (a) dramatic shallowing in water depth, and/or (b) decline in nutrient availability in surface waters. Shallowing over the study area through the Late Opoitian to Recent was progressive rather than sudden, which suggests that the often-dramatic fluctuations observed in planktic percentage were a response to changes in productivity rather than changes in depth, although paleodepth changes do overprint this signal. Given the absence of major zones of oceanic upwelling, changes in local current cells (coastal upwelling) may be invoked to explain the observations.

As there are no major oceanic fronts in the vicinity of the study area, the most likely source of a steady nutrient supply was via the Tasman Current. The nutrient load is derived from the inherent nutrient loading of the watermasses transported by the current (e.g., mode water) with additional minerals (dissolved clays) originating from erosion of the Southern Alps (Robert et al., 1986). The model proposes that through the Plio-Pleistocene, progradation of the continental margin may have acted to deflect the northern arm of the Tasman Current further offshore, particularly during periods of low sea level. As a consequence of lowered sea level, any localised coastal upwelling may also have been affected, and this could account for or accentuate the effects of changing nutrient supply to the planktic foraminifera in the region.

To summarise, lower (relative) sea level conditions, using planktic and benthic foraminiferal data as proxies, may be identified on the basis of several key criteria:

1. benthic diversity peaks, coincident with textural spikes and a large contribution of shallow water fauna in an essentially deep water assemblage;

2. dominance of opportunistic benthic species in a sample interval, particularly when this occurs immediately above, or coincident with, a textural and benthic diversity spike;
3. a positive correlation between lower planktic percentages, and peaks in benthic diversity and/or opportunistic benthic species spikes.

A simplistic model for sea level change on the basis of foraminiferal data is illustrated in Fig. 6.21. Cyclical trends are integrated with trends interpreted from wireline and seismic data to generate a preliminary model for the identification of sequences through the Giant Foresets Formation (following section).

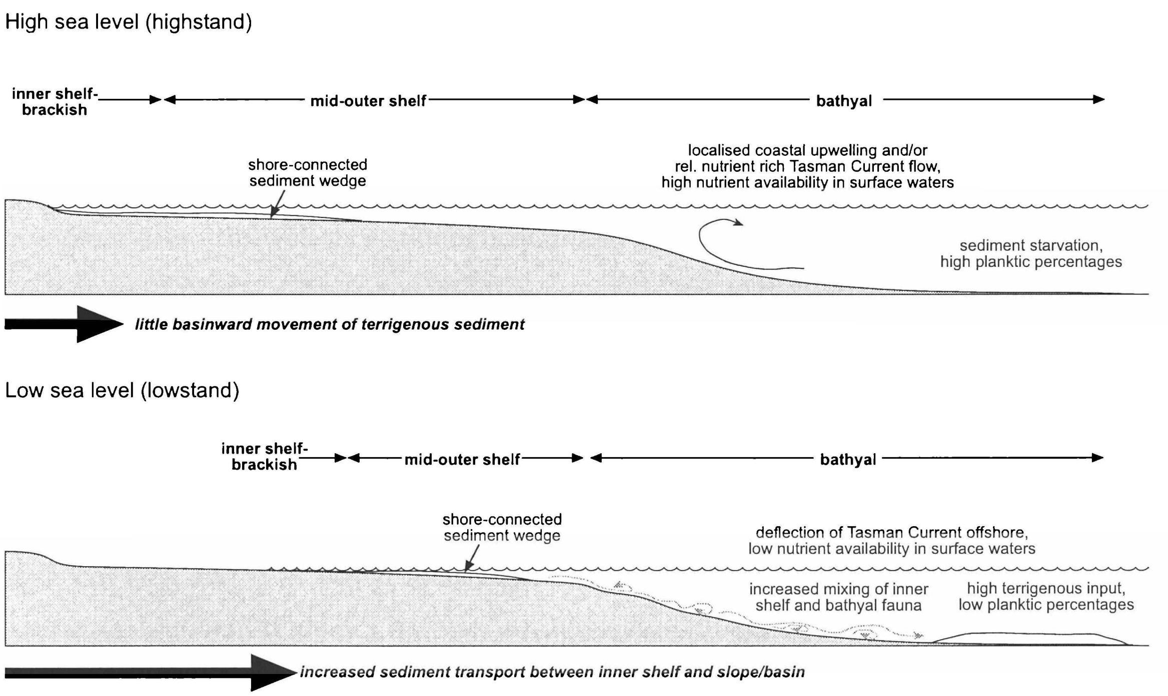


Fig. 6.21: Schematic diagram illustrating the association between sea level and observed foraminiferal trends. The model predicts that (a) lowered sea level alters the local coastal current system, resulting in reduced coastal upwelling, and possibly deflecting the more nutrient-rich Tasman Current away from the study area, thus decreasing the productivity of planktic foraminifera in the surface waters; and (b) closer proximity to the paleoshoreline increases the efficiency of transport of sediment from the inner and mid shelf to the slope and basin floor, resulting in increased mixing of shallow and deep water benthic faunas, and increased (catastrophic) nutrient flux to the sea floor.

6.5 Integrated cyclicity – a preliminary model

This section builds upon the cyclical signatures identified in Chapters 4 and 5, and integrate these with the model outlined above. The Giant Foresets Formation represents the latest phase of a 2rd-order regressional cycle (Rangitikei Megasequence of Kamp et al., 2002), which has resulted in the progradation and aggradation of the continental shelf and shelf margin. While 4^{th}- and possibly 5^{th}-order cyclicity has been interpreted from wireline log motifs, and some seismic horizons/sequence boundaries have been identified, a model for consistently identifying cycles within the Giant Foresets Formation using these methodologies is at this stage is not realistic. However, by combining seismic and wireline interpretations with foraminiferal data, a more robust method for identifying cyclical sea level change can be preliminarily applied to the Giant Foresets Formation sedimentary succession.

Enclosure 4 summarises the wireline, seismic and foraminiferal signatures interpreted for Arawa-1, Ariki-1, Kora-1, Wainui-1, and Mangaa-1. This has allowed the identification of a number of probable sequences in each of these well sections, some of which can be correlated between well sections. Each sequence is delineated on the basis of several of the following criteria:

1. a wireline motif characterised by a succession of overall upward-fining or upward-coarsening units, and identification of slumped units;
2. 'kick' points at the base of each of these wireline motifs;
3. correlation between low planktic percentages, an increase in shallow water benthic taxa, and a textural spike (seen either on the textural curve or wireline logs);
4. opportunistic benthic spike co-incident with or immediately after a sudden influx of shallow water taxa;
5. a seismic sequence boundary where identified from Chapter 5;

The best sequence cyclicity, as discussed in Chapter 4, is observed in the Mangaa-1 well section, expressed by the well-developed-blocky wireline motif of the Mangaa Formation. Numerous sandstone-siltstone couplets, possible 5^{th}-order sequences, are identified in the Opoitian and (particularly) the Waipipian Stage. In the upper part of the well section, cyclical trends are much less obvious, however, several distinct kicks on the SP log, some of which correlate to seismic unit boundaries, are prominent. The other well sections on Enclosure 4 do not display the same degree of cyclicity for the Opoitian and Waipipian Stages as seen in Mangaa-1, although several sequences (4^{th}-order) are identified.

Ariki-1 displays good correlation between interpreted seismic sequence boundaries and the occurrence of shallow water and opportunistic benthic species spikes. These boundaries often show a relationship with lower planktic percentages also. Sequences are more ambiguous in Kora-1, particularly through the middle (Early Nukumaruan) part of the well section where fauna are scarce. Below this barren interval, sequences are identified on the basis of benthic (shallow and opportunistic) abundance spikes and lower planktic percentages. Several of these boundaries correlate to seismic unit boundaries. Above the barren interval, benthic spikes are much more sporadic, and sequences are best delineated on the basis of wireline log motif and seismic unit boundaries. In this interval, several postulated boundaries correlate to a shallowing trend on the paleobathymetric curve.

Several of the seismic sequences identified in Chapter 5 have a thin shelfal component at Arawa-1 and Wainui-1, although these units are above the interval sampled for foraminiferal analysis. In the Wainui-1 well section, shallow water benthic spikes often occur independent of opportunistic species spikes, possibly suggesting that sediment influxes from the shelf were not great enough to smother existing communities and that often sediment was sourced from the upper slope to outer shelf, rather than inner shelf. Where shallow water benthic spikes do occur, these often correlate well to distinct kicks on the gamma-ray log, followed by upward-fining wireline motif, and in several instances, to seismic sequence boundaries. Sequences are much more difficult to identify in Opoitian to Mangapanian strata than for Nukumaruan and Castlecliffian strata.

Up to nine cycles can be identified at Arawa-1, although none of them (except seismic units/ sequences A54 and A73 which occur above the sampled interval in each well section) can be directly correlated to Wainui-1 through seismic reflection profiles. As with other well sections, sequences at Arawa-1 display 4^{th}-order periodicity.

While the integration of these methodologies has identified a number of sequences, it is obvious that there are significant problems with the consistent identification of sequences, particularly when compared with the Wanganui Basin record sequences, which are based on outcrop examinations.

- The composite nature and comparatively wide spacing of cuttings samples means that resolution in boreholes is not adequate enough to pick up all environmental signals through the sedimentary section. This is compounded by samples that have very low specimen counts, thus being discounted from any analysis.

- In deeper water situations, eustatic sea level change is not in itself enough to change bathyal-restricted habitats of benthic foraminifera. On the shelf however, a fall or rise in sea level of c.60-120 m will severely alter the dynamics of shallow water environments. Hence, when a site is situated at greater than shelf depths, few, and then only slight, changes in paleobathymetry are noted.
- Most clastic material supplied to the foresets is presumed to have done so during lower sea level conditions, and very little during high sea level conditions and thus an almost continuous signal of inner to mid shelf faunas may be preserved.
- Geophysical wireline logs are extremely useful in helping to identify cyclical trends and sequences. These trends are best developed through very sandy successions (e.g., sandstone-mudstone couplets of the Mangaa Formation) and shelfal successions (e.g., upper part of Giant Foresets Formation). Thus the muted wireline response that much of the Giant Foresets Formation sedimentary section displays due to lack of lithological variation is not conducive to consistently identifying sequences within a well section, much less correlation between well sections, as evidenced in Enclosure 4. Wide spacing of well sites contributes to correlation difficulties.
- Sequences are probably best defined by seismic criteria, such as achieved for seismic reflection profile P95-103. Because of the characteristics of the foresets (channel-levee complexes and slope fans rather than regressive and lowstand sandstone-dominated basin floor fans, lack of onlap, and prevalence of slumping), the approach taken by this study (i.e., identification of architectural elements) is considered preferable to the Vail approach to sequence identification. This however needs to be developed much further and applied to several seismic sections.
- Industry acquired seismic does not have the resolution required to identify the 5^{th}- and 6^{th}-order cycles that are identified in outcrop in Wanganui Basin for the same age strata. Thus the highest order cyclicity that can realistically be observed on seismic reflection profiles is 4^{th}-order (c. 400 ka).

The Giant Foresets Formation is an obviously cyclical succession, and in several well sections, upwards of fourteen 4^{th}-order cycles have been delineated. However, much more work (probably supported by a continuous coring programme) needs to be undertaken to build a robust and reliable model that can be transferred consistently from well section to well section.

6.6 Conclusions

- Detailed studies of benthic and planktic foraminifera are essential in reconstructing paleoenvironmental trends through time, and establishing paleobathymetry and watermass conditions.
- Paleobathymetric curves for Arawa-1, Ariki-1, Kora-1, Mangaa-1, and Wainui-1 show a progressive shallowing from the Late Miocene to Pleistocene, with all sites reaching shelfal depths by the Late Nukumaruan to Castlecliffian.
- A sharp decrease in planktic percentage and an apparently dramatic change in watermass conditions is observed between latest Miocene/Opoitian strata and Waipipian strata. Paleobathymetric curves indicate that water depths did not change as significantly as changing watermass conditions would imply. Thus the initially dramatic change may be related to deflection of local current systems away from northern Taranaki Basin by the advancing progradational front, affecting surface water nutrient supply and therefore planktic percentage. Increased suspended sediment load associated with increasing sediment flux through the Late Pliocene and Pleistocene could account for an observed planktic increase during this interval at Ariki-1.
- Benthic foraminiferal data are particularly useful in demonstrating sediment pathways across the study area; benthic assemblages, even at the deepest sites, are often characterised by the postmortem mixing of shallow water faunas, indicating that at times there was effective sediment transport across the shelf to the slope and basin floor.
- The integration of benthic foraminiferal data, including faunal composition, diversity indices, and species dominance, with watermass conditions and textural curves, can provide a proxy for indicating periods of lowered relative sea level in the absence of other criteria. However, inherent difficulties with cuttings samples and sample resolution problems mean that not all paleoenvironmental signals will be identified.
- The integration of seismic, wireline and paleoenvironmental data has enabled the recognition of ten or more 4^{th}-order sequences in northern Taranaki Basin. However, correlation of sequences, predominantly because of the wide spacing of wells, is in reality only possible using seismic reflection profiles. More detailed research on this aspect of the Giant Foresets Formation needs to be undertaken to construct a more robust sequence stratigraphic model.

Chapter 7: Dynamics and evolution of a prograding continental margin

Chapter 7: Dynamics and evolution of a prograding continental margin

7.1 Introduction

The purpose of this chapter is twofold; (1) to execute and analyse the results of backstripping and decompaction of each of the eleven well sections utilised in this study, as well as to backstrip and decompact a complete seismic reflection profile for palinspastic restoration; and (2), to collate relevant information from this and previous chapters into a synthesis, incorporating a series of paleogeographic maps, for the period leading up to, and during deposition of the Giant Foresets Formation. A third part to this chapter involves a brief outline of the petroleum systems of the northern Taranaki Basin, and discussion about the impact the Giant Foresets Formation has had on the thermal regime and petroleum generation of northern Taranaki Basin.

7.2 Methodology

One and two-dimensional backstripping and decompaction of sedimentary units (geohistory analysis) is an effective way of summarising the geological evolution and burial history of an exploration well section or of a cross section. This is achieved by mathematically removing layers sequentially, thereby allowing underlying units to rebound back to the original surface while concurrently being decompacted. By progressively removing the effects of sediment loading (including compaction), eustasy and paleoenvironments (Pekar et al., 2000) from basin subsidence, a tectonic subsidence curve can be obtained. Derivative data about sediment accumulation rates, the identification of episodes of tectonic uplift and/or subsidence, and the thermal maturation history of source rocks can also be interpreted. While the calculations involved in backstripping may be done by hand, typically they are performed through computer programmes which integrate lithostratigraphic, biostratigraphic, and paleobathymetric information in routines that compute decompacted sediment thickness and subsidence/uplift as a function of time (Hayward, 1987; Sircombe, 1993). The programmes used in this study are described below.

7.2.1 Geohistory analysis

While geohistory plots have previously been generated for most well sections in the study area (e.g., Hayward, 1987; Hayward and Wood, 1989; various well completion reports), it was decided to rerun all analyses for each of the wells used in this study using newly acquired or revised information. The input parameters (discussed below) and graphical results are presented in Appendix 7 and discussed in section 7.3.

Geohistory analysis was carried out using the programme GeoHist+, a programme developed by K. Sircombe (Sircombe, 1993). GeoHist+ is a modification of the programme GEO_HIST, developed by Wood (1989) and based on the decompaction technique of Falvey and Deighton (1982). The theory and systematics behind the usage of GeoHist+ are only briefly described here. For a full discussion on the development, programme syntax, and attributes of this programme, refer to Sircombe (1993).

Four input parameters are required for backstripping and decompaction:

1. compacted thickness of units (depth to base minus depth to top of unit);
2. water depth at the time of deposition (paleobathymetry);
3. age (in millions of years) at the base of the unit; and
4. the relative percentages of each lithology making up the unit.

Units were delineated based on major lithological and/or biostratigraphical criteria. Up to twenty stratigraphic units can be delineated for each well section. Units are systematically decompacted (i.e., the present day thickness is corrected to account for the progressive loss of porosity with depth of burial; Sircombe, 1993) during the process of generating a geohistory plot. GeoHist+ also accounts for paleo sea level, based on the long-term curves of Haq et al. (1987). Subsidence driven by the effect of the water column is included in the GeoHist+ calculations of the tectonic curve. Where a unit is described as a mixture of lithologies, GeoHist+ calculates each component automatically as a weighted average. Once the programme (and the operator) has checked the data for irregularities, the units are sequentially removed from the top of the sedimentary column, with remaining units decompacted and adjusted for paleobathymetry and eustasy, and a tectonic subsidence curve calculated (Sircombe, 1993). The subsidence curve records the subsidence and uplift history for the top of the basement, for the boundaries of each unit, and for the present-day seabed. Unconformities and erosional events can also be incorporated into the calculations using GeoHist+. For a hiatus, the depth to base is entered as equal to the depth to top (i.e., zero thickness) for the appropriate time span. For an erosional event, the depth to base is less than the depth to top (i.e., a negative thickness).

7.2.2 Palinspastic restoration

Backstripping and decompaction is normally performed on a single well section or on a series of well sections. For this study, a two-dimensional backstripping programme (excel-based macro), written by K. Sircombe (1993) was run on a complete (post Miocene) seismic reflection profile (line P95-158). Input parameters are similar to those outlined above, with paleobathymetry, lithology and thickness of each interpreted *seismic* unit included. Paleobathymetry was established from foraminiferal data for the wells Tangaroa-1 and Kahawai-1. Depth information from Mangaa-1 and Awatea-1 (tied into line P95-158 via lines P95-138 and P95-115) were included where appropriate. All depths in TWT for seismic units were converted to depths in metres using the binomial supplied by Geosphere Exploration Services Ltd. and presented in Chapter 5. Lithologies were established from wireline and lithological data. However, rather than using the relative percentages of lithological components, the combined porosity and compaction coefficient for each seismic unit were used instead.

Porosities were determined by estimating the relative proportions of sand, silt, mud, and limestone/marl of each individual seismic unit through which a borehole was drilled. Using initial (surface) porosity values and standard compaction coefficients (after Armstrong et al., 1998; Table 7.1), porosity with depth was calculated using Equation 1.

$$\text{porosity} = po \exp(-z/d) \qquad \textit{Equation 1}$$

where po = initial porosity at the surface, z = depth, and d = compaction coefficient.

The values obtained for each seismic unit were then combined using a mixing law (Equation 2) to provide porosity variation with depth.

$$\text{combined porosity } (1/p) = \text{summation } (vi/pi) \qquad \textit{Equation 2}$$

where p = porosity, vi = proportion of a particular lithology, and pi = porosity determined from Equation 1.

Table 7.1: Initial porosity and compaction coefficient values (from Armstrong et al., 1998).

	Sand	Silt	Mud	Limestone
Initial porosity	45	49	54	70
Compaction coefficient	3000	2500	2000	1400

Similarly, decompaction coefficients (after Armstrong et al., 1998) were mixed according to the proportion of each lithology, to obtain a value representative of the mixed porosity.

Line P95-158 is approximately 68 km long, and has only two boreholes drilled along its length – Tangaroa-1 at the western end, and Kahawai-1 at the eastern end. While lithological variations along the length of this seismic line are apparent (Fig. 5.28), the change in paleobathymetric depth for a single seismic unit is marked. For this reason, the seismic reflection profile was broken into six parts; paleobathymetry and lithologies (and thus combined porosities and compaction coefficients) were estimated using position on the shelf-slope-basin floor profile and information from drillholes (Tangaroa-1, Kahawai-1, Awatea-1, and Mangaa-1). The backstripping macro was then run on each of the six parts. As with GeoHist+, as each layer (in this case, seismic unit) was stripped off, the remaining layers rebounded to their pre-compacted configuration. Graphs were created in Microsoft Excel, exported to Macromedia Freehand 10™, and rejoined. The resultant figure is discussed in section 7.3.4.

7.2.3 Sources of error

Backstripping and decompaction programmes, such as the ones utilised in this study, are inherently subject to several errors. These are summarised in Hayward (1987) and Hayward and Wood (1989), and are briefly discussed here.

Sources of inaccuracies arise from all input parameters. Paleodepths (and relative ages) are based on foraminiferal data derived from well reports. Only the four wells assessed for paleoenvironmental information (Arawa-1, Ariki-1, Kora-1 and Wainui-1, Chapter 6) plus Mangaa-1 have paleodepth information of any detail through the Giant Foresets Formation. For all other wells, Plio-Pleistocene information is much less detailed. However, while it is difficult to establish paleodepths with absolute certainty, the accuracy of paleodepth estimates for Neogene and Paleogene sediments are determined to be more accurate than for Cretaceous sediments (Hayward and Wood, 1989). For all layers backstripped using the GeoHist+ programme, where a range of depths were possible, depths from the mid to higher (deeper) end of the depth range were assigned. These were checked against depths assigned by Hayward and Wood (1989), with similar depths being assigned for wells in common. The perceived accuracy of assigned paleodepths from Hayward and Wood (1989) are listed below:

depth range estimate	***accuracy***
0-100 m	±30 m
100-200 m	±50 m
200-400	±100 m
400-1000	±200-300 m
1000-2000	±500 m

Thus successively deeper stratigraphic units have greater degree error estimates associated with paleodepth calculations. The error may be even greater for line P95-158. This is because of the limited amount of depth information that can be extrapolated from the wells lying on or near line P95-158 (i.e., Tangaroa-1, Kahawai-1, Awatea-1 and Mangaa-1). Using these wells, and interpolating depths relative to the paleoshelf break where necessary, a single (average) depth was assigned for each of the six segments. Invariably, depths change along the length of each segment, thus the chance for paleodepth errors is increased. This also holds true for lithologies, although the errors are probably not as great.

Layers sequentially backstripped using GeoHist+ were often delineated on the basis of biostratigraphic stages (New Zealand). Absolute ages were determined from the interim New Zealand geological time scale of Morgans et al. (1997). Ages of seismic units, where they fell between stages, were interpolated from stage boundary ages. Hayward and Wood (1989) suggest that dating inaccuracies should be of no real significance, except where an estimate of the age at the base of a sequence was made because the drillhole did not reach basement. Where the drillhole has not reached basement, the full thickness of the Cretaceous-Cenozoic sequence has been taken from Hayward and Wood (1989). For wells not evaluated by Hayward and Wood (1989), an approximate estimate of the thickness to basement (if not drilled) was made from structure contour maps of Thrasher and Cahill (1990). Thus the thickness and resultant compaction calculations from TD to top basement are a source of inaccuracy.

Several other sources of inaccuracy are discussed in Hayward and Wood (1989). These include the compaction formulae, the porosity and percentage of each lithological component within a unit, difficulties in allowing for undercompaction due to overpressuring, and the assumption that compaction and subsidence due to loading are assumed to occur almost instantaneously as burial progresses. Waltham et al. (2000) suggests that error analysis of subsidence curves should form part of geohistory analyses, but also concede that this is a complex problem because of the iterative and nonlinear nature of the decompaction calculations. There is limited opportunity to illustrate the uncertainties on the graphical outputs. These uncertainties need to be qualitatively factored into calculations.

7.3 Geohistory analysis - results

A geohistory plot has been generated for each well. For each plot three curves are shown, each of which summarises an aspect of the geohistory of the sequence at that site (Hayward, 1987). These are the paleo-sea level curve (based on Haq et al. 1987), the total subsidence curve, and the tectonic subsidence curve. A fourth curve, paleo-sea bottom (obtained from paleodepths interpreted in Chapter 6) is also plotted. Each well section is presented in Fig. 7.1 and Appendix 7.

7.3.1 Tectonic history of wells in northern Taranaki Basin

Several characteristics of the geohistory curves generated by this and earlier studies are immediately obvious on their perusal. Most apparent is a Cenozoic interval of sustained subsidence and concomitant increase in water depth, from paralic depths at the end of the Cretaceous, to lower bathyal water depths of 800 to 1500 m during the Early to Middle Miocene. A second obvious characteristic of all curves (with the exception of Turi-1) is the rapid shallowing from bathyal to shelf depths through the Late Miocene and into the Pliocene. For a more detailed discussion of the pre-Miocene history expressed in Taranaki Basin wells, see Hayward (1987) and Holt and Stern (1991). The rest of this section emphasises the Late Miocene – Present geohistory of northern Taranaki Basin wells.

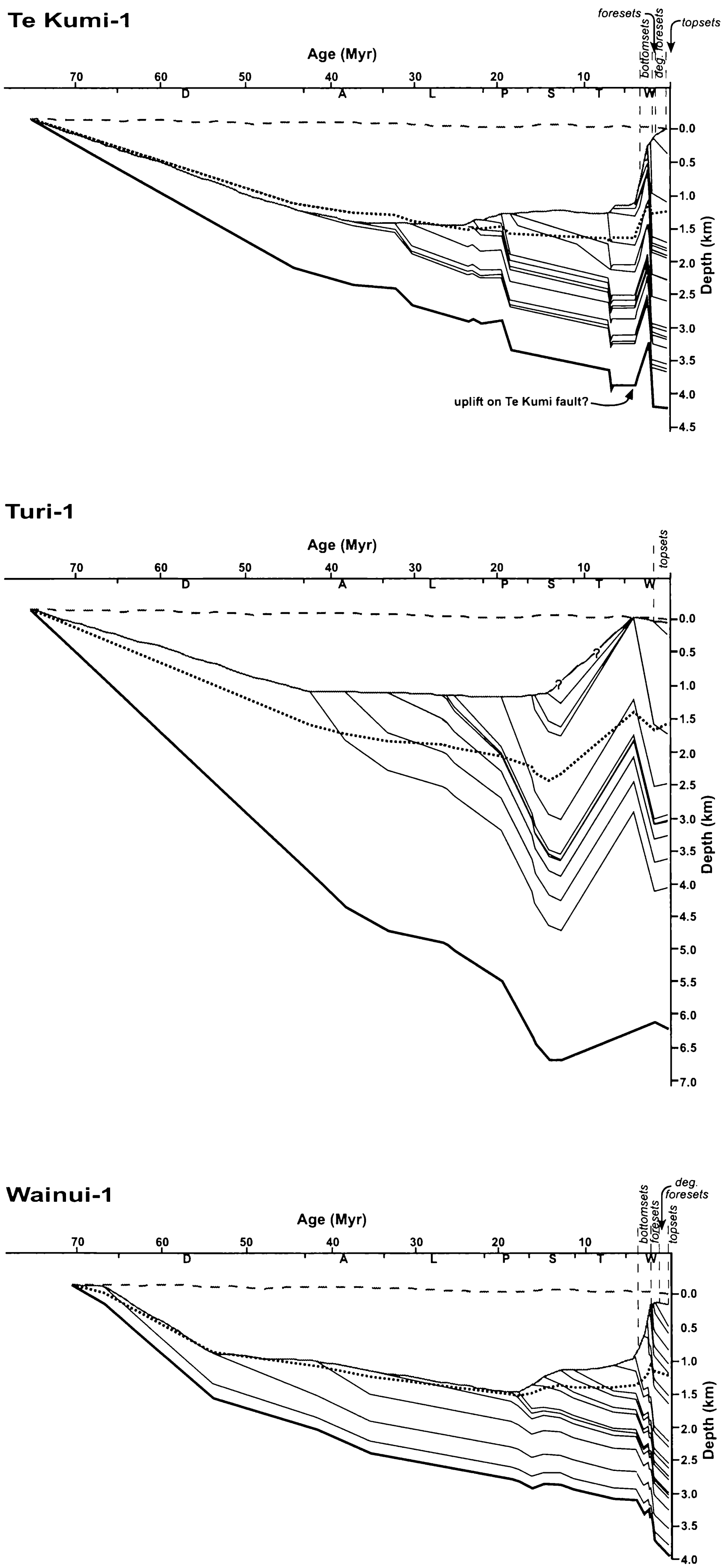


Fig. 7.1 continued.

At most well sites, the tectonic curve displays staggered uplift during the Pliocene or Early Pleistocene, even on the Western Stable Platform. This was also noticed by Hayward and Wood (1989), who suggest that, particularly on the Western Stable Platform, which has been relatively tectonically quiescent since the Oligocene, the uplift of the tectonic curve is an artifact of the assumptions made in the formulae used in the calculations. In particular, they suggest that the artifacts arise from the crust not having sufficient time after deposition of the succession to adjust following the rapid loading of the Giant Foresets Formation. Alternatively they also suggested that the pre-foresets succession might be undercompacted.

At some sites a sharp peak on both the tectonic and total subsidence curves does however indicate that some uplift occurred during the Late Miocene to Early Pliocene, presumed to be related to fault movement. These peaks are pronounced at Kahawai-1, Te Kumi-1, and Turi-1 (note that the amount of uplift on Turi-1 is slightly underestimated due to limitations of the geohistory programme). At Turi-1, approximately 1200 m of erosion is estimated (Armstrong et al., 1998), encompassing much of the Late Miocene to Pleistocene sedimentary section. This suggests that the region in the vicinity of Turi-1 was emergent for a period. Increasing erosion across the Turi Fault Zone is inferred by Armstrong et al. (1998), who indicate that up to 1800 m of Miocene to Late Pliocene strata may have been eroded from the northeastern region of Taranaki Basin.

The Pliocene peaks noted at Kahawai-1 and Te Kumi-1 correspond to fault-related movement in the immediate vicinity of each well site. At Te Kumi-1, an Early Pliocene peak (possibly slightly exaggerated due to the loading effect?) and associated onlap of Late Opoitian to Waipipian strata suggest that uplift on Te Kumi Fault during the Late Miocene to Pliocene created a localised paleogeographic high. This may have contributed to condensation and development of the Ariki Formation in this region, as well as helping to create the western edge of the Northern Graben. At Kahawai-1, the tectonic curve suggests that there were two periods of movement, a more sustained one during the Middle Miocene, and another sharper peak during the Opoitian to Waipipian. Vertical offset, in the order of 200-300 m, of seismic units A16-A39 at Kahawai-1 is consistent with Opoitian-Mangapanian uplift.

Subsidence associated with formation of the Northern Graben is illustrated as a relatively sharp increase in total subsidence in the mid to Late Tongaporutuan, in response to the development of extension. This created a domed structure (roll over anticline?) to the west of the graben, on to which the Ariki Formation onlaps. While most of the Western Stable Platform was gently subsiding during the Late Miocene and Pliocene, subsidence of the Northern Graben is evident in both the total subsidence and tectonic subsidence curves, and the bathymetry (paleoseabottom) curve for Mangaa-1 and Awatea-1. At both sites, paleodepths deepened by approximately 300 m.

The marked shallowing from bathyal to shelfal depths for most northern Taranaki Basin well sites began in the Kapitean. This compares with a middle Miocene start to regression in southern Taranaki Basin (King and Thrasher, 1996). Most sites shallowed from a maximum water depth of 800-1500 m to 100-200 m or less during the Kapitean to Nukumaruan Stages, beginning earlier in the south than the north. Locally (e.g., Kahawai-1) uplift contributed to this shallowing. Initial shallowing occurred through aggradation of bottomset (basin floor) units, including Manganui, Mohakatino and Mangaa Formations, as well as bottomsets of the Giant Foresets Formation. At most sites, the steepest part of the paleoseabottom curve occurs during the Plio-Pleistocene, and is thus related to deposition of the Mangaa Formation (Northern Graben) and Giant Foresets Formation. While one would intuitively think that most of the shallowing would occur during deposition of the foreset strata, in fact, it is during deposition of bottomset strata that the region shallowed most dramatically, with water depths decreasing from middle to lower bathyal (600 – 1500 m+) to upper and middle bathyal depths (200 – 600 m+).

While much of the shallowing of Northern Taranaki Basin occurred through deposition of bottomset strata, invariably a sharp increase in the total subsidence curve coincides with the appearance of the prograding wedge (foresets) at each site. While there was a further 500-1500 m of total subsidence at each site during the Plio-Pleistocene as a result of sediment loading, sedimentation rates outpaced subsidence rates so that the net effect was a continuation of shallowing water depths. As the progradational wedge moved past each site, outer shelf depths were achieved (corresponding to topset beds). All topset sequences were deposited at shelf depths (less than 200 m).

7.3.2 Plio-Pleistocene sedimentation rates

Pliocene to Pleistocene accumulation rates (decompacted) for each of the seismic divisions (bottomsets, foresets, degradational foresets and topsets) have been calculated from geohistory plots for each well. These are presented in Table 7.2 and illustrated in Fig. 7.2. Several trends are immediately obvious:

1. foresets (slope facies) invariably have the highest sedimentation rates, followed by degradational foresets;
2. topsets (shelf facies) are generally associated with the lowest sedimentation rates;
3. sedimentation rates for foresets generally increase to the north and west.

Table 7.2: Decompacted sedimentation rates (m/1000 years) for Plio-Pleistocene strata, northern Taranaki Basin.

Well	Bottomsets	Foresets	Degradational foresets	Topsets
Arawa-1		0.32		0.42
Ariki-1	0.35	3.4	0.65	0.12
Awatea-1	0.58	1.4		0.36
Kahawai-1	0.41	0.90		0.50
Kora-1	0.86	0.54		0.47
Mangaa-1	0.60	2.8		0.38
Taimana-1		0.41		0.27
Tangaroa-1	0.48	4.3	0.98	0.09
Te Kumi-1	0.50	6.08	0.56	0.17
Turi-1				0.11
Wainui-1	0.47	1.60	0.20	0.12

Rates of deposition for basin floor facies (bottomsets) are generally fairly consistent, averaging around 0.53 m/1000 yrs. This is interpreted to reflect the limited amount of terrigenous sediment reaching the northern parts of the study area during the Late Miocene to Early Pliocene. The slow sedimentation rates obtained for foresets facies for Arawa-1 and Taimana-1 (< 0.5 m/1000 yrs.) reflect the initially slow advancement of the foreset front into the southern part of the study area. Through the Waipipian to Nukumaruan Stages the foreset front advanced at an increasingly fast rate through the central and northern parts of the study area, and this is reflected in increasingly high sedimentation rates for foreset strata. The anomalously high sedimentation rates calculated for Ariki-1, Tangaroa-1, and Te Kumi-1 are inflated due to loss of foreset strata through significant incision of overlying degradational foreset units, and in reality probably approximate the rates obtained for Wainui-1, Mangaa-1 and Awatea-1.

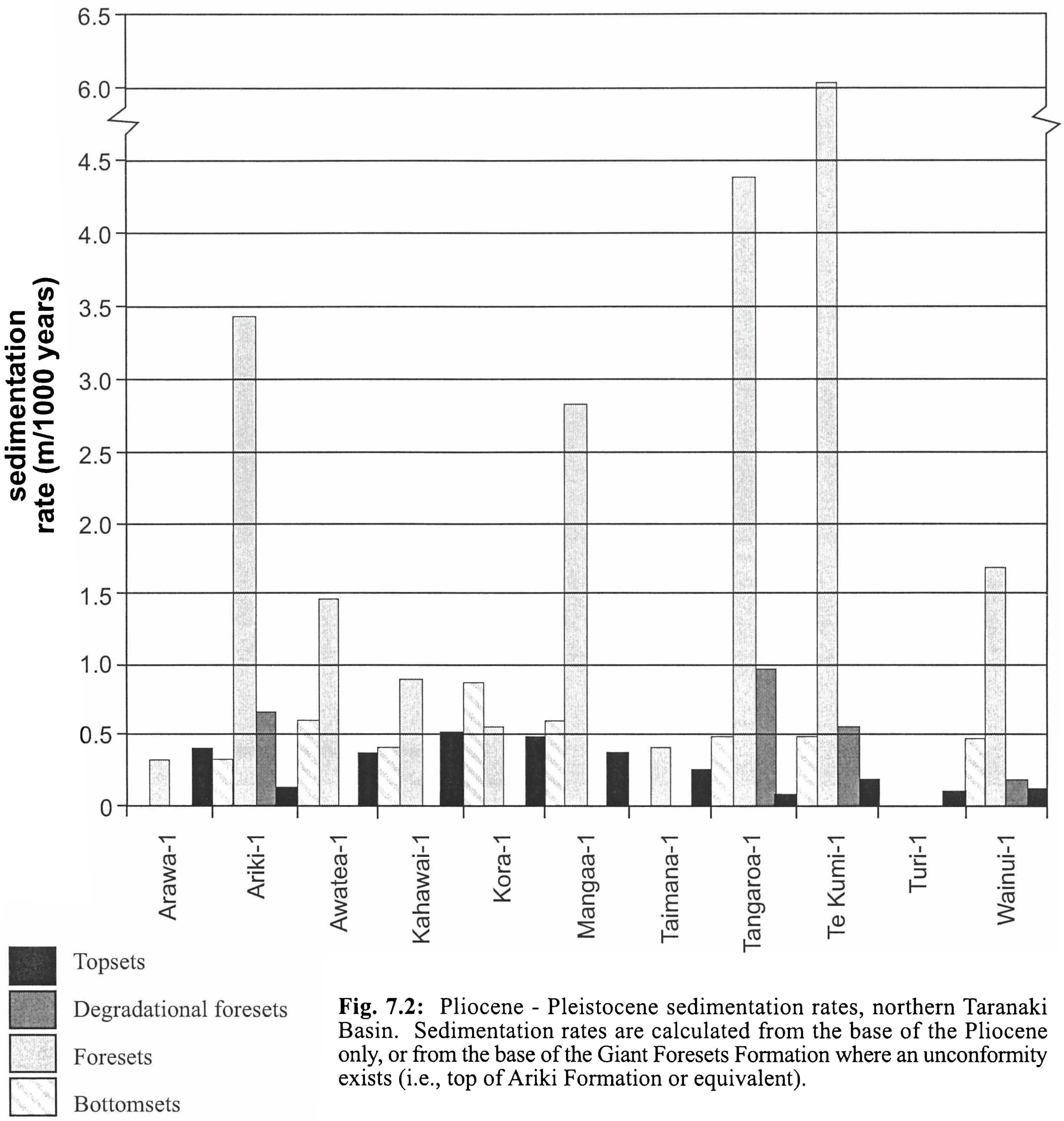
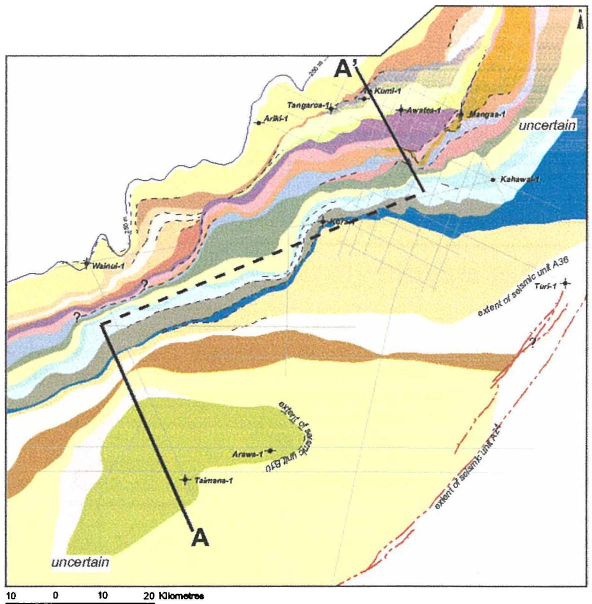
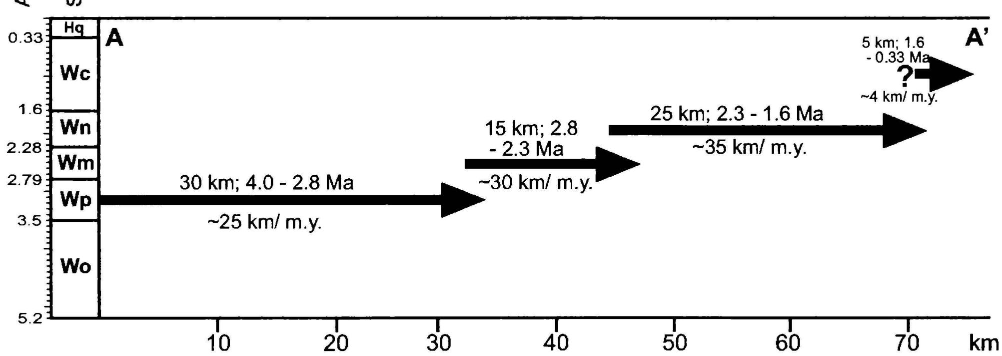


Fig. 7.2: Pliocene - Pleistocene sedimentation rates, northern Taranaki Basin. Sedimentation rates are calculated from the base of the Pliocene only, or from the base of the Giant Foresets Formation where an unconformity exists (i.e., top of Ariki Formation or equivalent).

The increasing rate of progradation of the foreset front through northern Taranaki Basin is also schematically illustrated in Fig. 7.3. This figure includes a transect across the study area, and relates the position of interpreted clinoform breaks to time and rate of advancement. Rates of progradation across northern Taranaki Basin increased sequentially from the Waipipian to Nukumaruan; the increasingly rapid advancement of the progradational front is related to an increasing volume of sediment derived from erosion of the Southern Alps. The sharp drop off in sedimentation rate during the Castlecliffian is probably related to the present highstand conditions.



(a) Line of transect across progradational shelf breaks (see Fig. 5.24 for legend).



(b) Distance and rate of progradation.

Fig. 7.3: Progradational rates through time. Note sequential increase in progradational rates into the Nukumaruan Stage, reflecting the acceleration of the advancement of the continental wedge across northern Taranaki Basin. New Zealand Stage Abbreviations - Wo, Opoitian; Wp, Waipipian; Wm, Mangapanian; Wn, Nukumaruan; Wc, Castlecliffian; Hq, Haweran.

Sediment accumulation rates are also high for degradational foresets (as high as 0.98 m/1000 yrs.), though much lower than for foreset facies. Degradational foreset strata are thickest at Ariki-1, Tangaroa-1, and Te Kumi-1 (Fig. 7.4), incising deeply into underlying progradational foreset strata at these sites. Topset strata record the slowest sedimentation rates (averaging approximately 0.27 m/1000 yrs.), indicating that sediment is bypassing the shelf and being deposited further basinward. The lowest rates are related to position relative to the modern day shelf break (excluding Turi-1). Ariki-1, Tangaroa-1, and Wainui-1, in particular, have only a very thin topset component that possibly only accumulated during the present interglacial highstand. The thickness of topsets (shelf facies) on the other hand is dependant on the elapsed time since the shelf edge prograded past a site. The more southern sites (Arawa-1, Taimana-1, Kora-1) and eastern sites (Kahawai-1) have a thicker topset component than more northerly sites (i.e., Ariki-1, Tangaroa-1, Te Kumi-1 and Wainui-1) (Fig. 7.4).

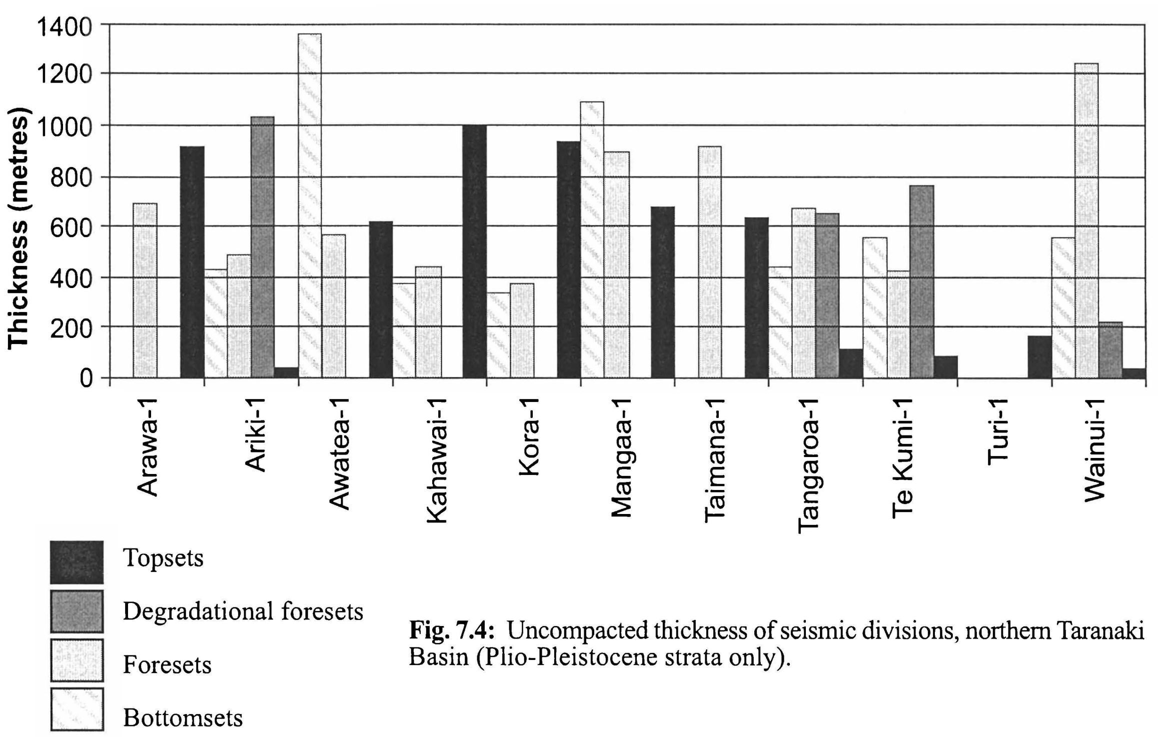


Fig. 7.4: Uncompacted thickness of seismic divisions, northern Taranaki Basin (Plio-Pleistocene strata only).

7.3.3 Palinspastic restoration of seismic reflection profile P95-158

Figure 7.5 shows the palinspastic restoration of seismic reflection profile P95-158. This figure displays (some of) the steps in backstripping and decompaction that sequentially restore the palinspastic position of depositional surfaces at intervals through the Late Opoitian to Recent.

The backstripping, decompaction, and palinspastic restoration of seismic line P95-158 illustrates the development of an asymmetrical Northern Graben. This development began in the Late Miocene, so that by the mid to Late Opoitian (Fig. 7.5a) the depression that formed as a result of extensional tectonics had developed enough that along the central axis of what was to become the Northern Graben was already becoming a focus for sediment deposition relative to the Western Stable Platform (Mangaa Formation). This invariably contributed to sediment starvation to the west of the graben. It is unclear how much movement took place on Kahawai Fault, but offset of seismic units A16 and A18 (Fig. 7.5a) suggest approximately 200 msecs of displacement occurred during the Opoitian. To the west, on the Western Stable Platform in the vicinity of Tangaroa-1, a low dome structure had begun to form, possibly a roll-over anticline in response to development of the Northern Graben.

Deepening of the graben continued into the Waipipian and Mangapanian (Fig. 7.5b), and the dome structure to the west of the graben became more pronounced. Another phase of displacement also occurred on Kahawai Fault, resulting in further vertical displacement between seismic units either side of the fault. The asymmetry of the Northern Graben became much more obvious at this time. Sediment accumulation continued during these stages, and by the end of the Waipipian, sediment accumulation had outpaced subsidence, and seismic units filled, and then spilled over the graben on to the Western Stable Platform. Water depths shallowed from approximately 1000 m, to less than 500 m during this period.

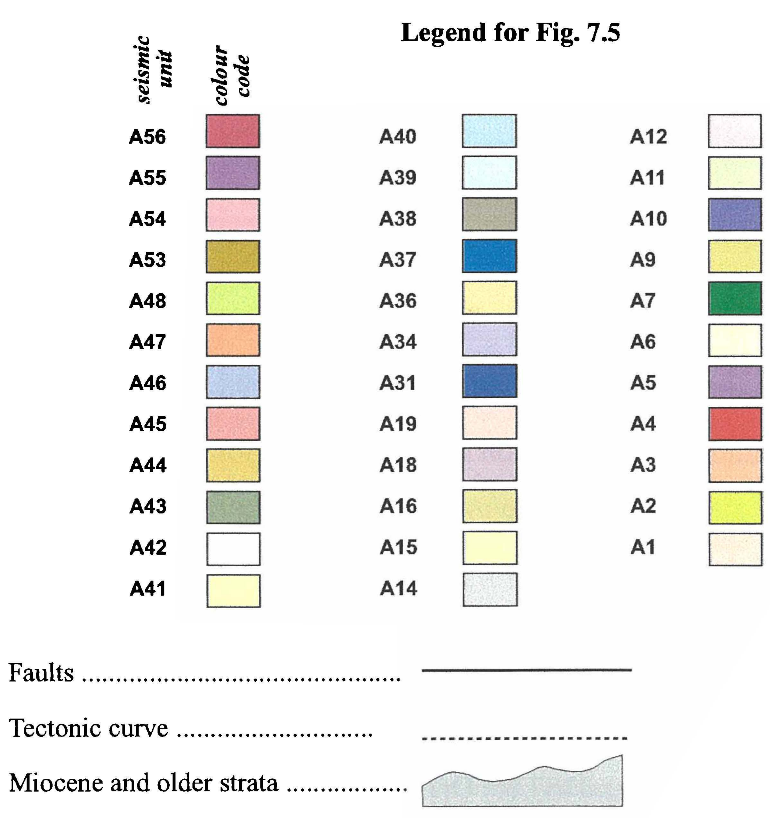


Fig. 7.5 (facing page): Palinspastic restoration of seismic reflection profile P95-158 for three intervals.

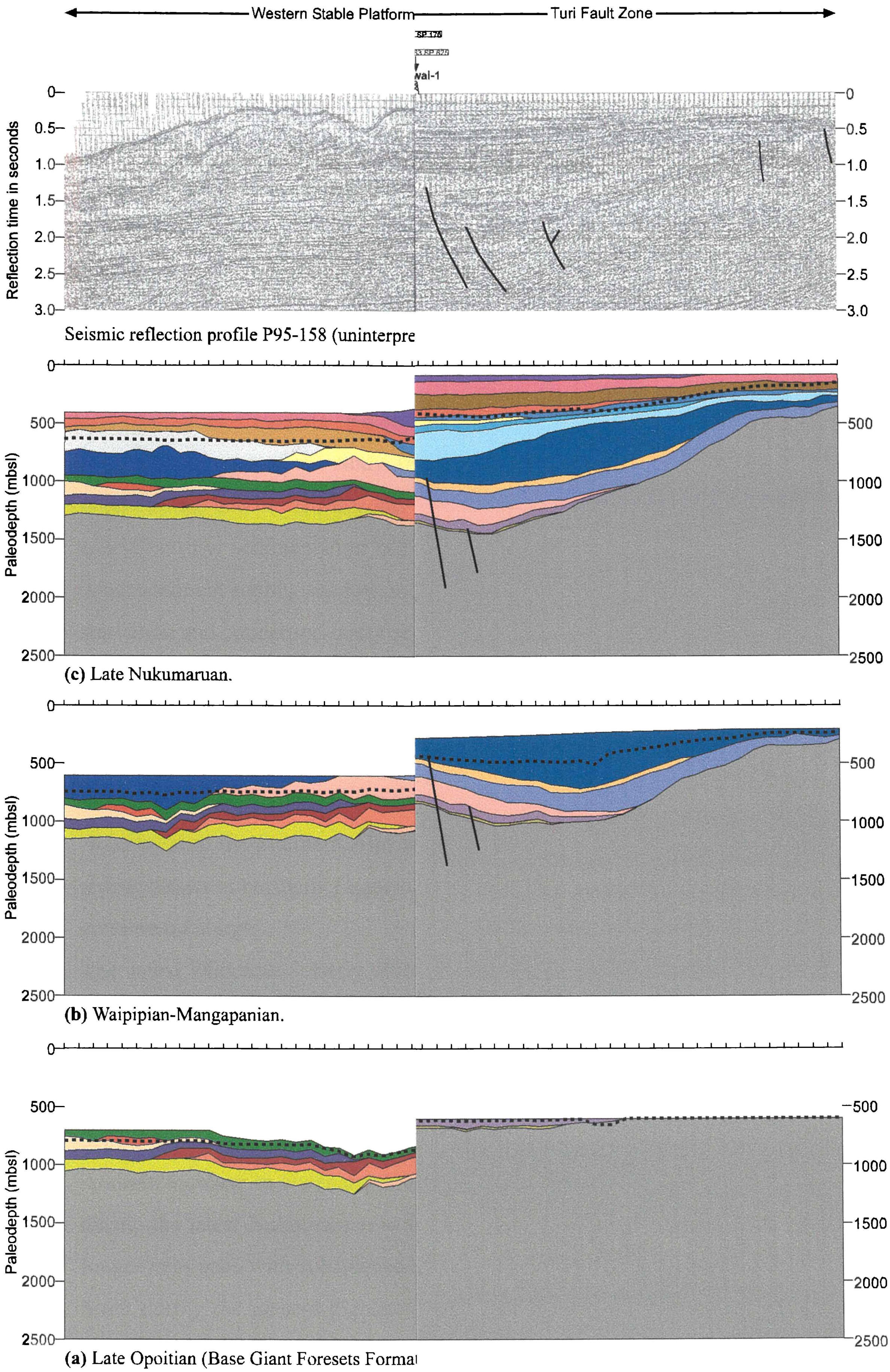


Fig. 7.5: Palinspastic restoration for three inte

Subsidence of the Northern Graben continued during the Nukumaruan (Fig. 7.5c), with the asymmetry of the graben possibly enhanced by the appearance of foreset strata prograding across the region. Geohistory plots for individual wells (Awatea-1 and Mangaa-1) suggest that 300-500 m of further subsidence can be attributed to the loading effect caused by foreset strata. However, sedimentation occurred rapidly enough that depositional rates outpaced subsidence rates, and water depths rapidly shallowed across the graben.

7.3.4 Summary of palinspastic restoration, backstripping and decompaction analysis

- Throughout most of the study area, subsidence had been relatively constant since the early part of the Cenozoic. However, by the Late Miocene, subsidence rates increased in the central part of the study area, and most sites were under their deepest water depths by the end of the Miocene.
- Late Cenozoic sedimentation rates in the study area have varied greatly. At times when sediment supply was low, much of the study area was characterised by sediment starvation and condensed marly sedimentation. This situation was accentuated on the Western Stable Platform by several factors, including subsidence of the Northern Graben and the capture of terrigenous sediment supply, and minor flexural doming (roll-over anticline?) west of the Northern Graben that created a paleohigh slightly elevated relative to surrounding areas.
- As erosion of the Southern Alps accelerated, voluminous sediment was available for redistribution in northern Taranaki Basin. This lead to northward progradation of the continental margin.
- The latest Miocene – Early Pliocene formation of the Northern Graben through extensional faulting created the initial depocentre that attracted sedimentation. Eventually however sedimentation exceeded subsidence, resulting in relatively rapid shallowing of the depositional surface from bathyal to outer shelf or shallower water depths during the Pliocene and Pleistocene.
- A slight increase in water depth noted at some wells, and the drop in sediment flux during the latest Nukumaruan to Castlecliffian, suggests that sedimentation rates no longer kept pace with subsidence rates. This could be attributed to opening of Cook Strait during the present highstand conditions, and capture of sediment in this tidal seaway.

7.4 Synthesis – late Neogene evolution of a prograding continental margin, northern Taranaki Basin

A main objective of this study was to better constrain the Late Miocene – Recent paleogeographic development of the northern part of Taranaki Basin, with emphasis on the Giant Foresets Formation. This has been achieved by integration of seismic reflection data, wireline log data, and foraminiferal paleoecologic data discussed in previous chapters, as well as utilizing the results of backstripping models as discussed above. The end result is a series of paleogeographic maps that illustrate and document the evolutionary development of the Giant Foresets Formation within the northern part of Taranaki Basin.

7.4.1 Early to Late Miocene (Figs. 7.6a,b, 7.7: c.23.8 - c.6 Ma)

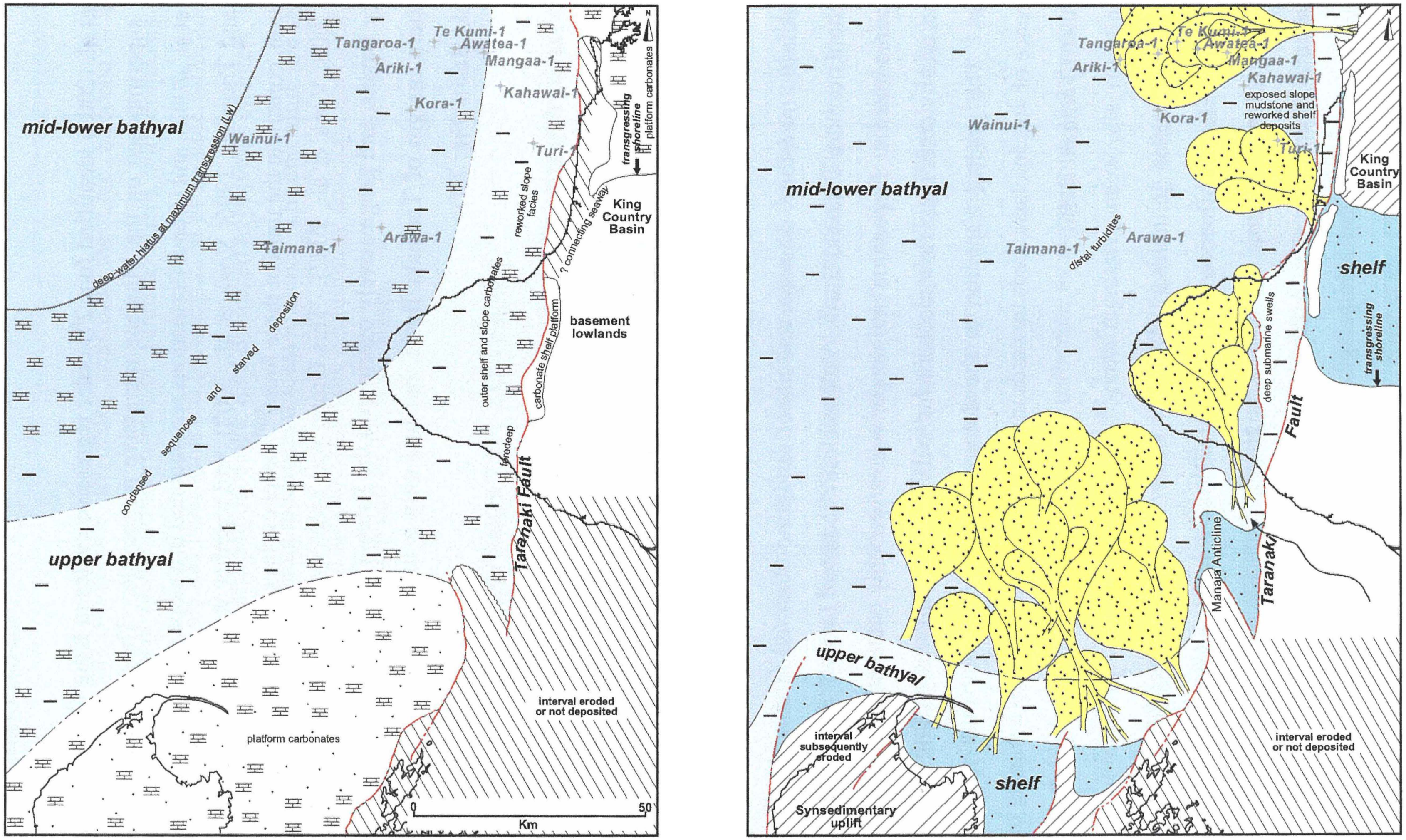
Rapid subsidence from the Late Oligocene marked a change in tectonic regime from a relatively quiescent subsidence phase, to one influenced by far-field effects of the evolving Australia-Pacific plate boundary (Bennett et al., 1992; King and Thrasher, 1992). The major structural expression of the development of a compressive regime across North Island was overthrusting of basement into Taranaki Basin along the Taranaki Fault during the Early Miocene. This truncated the sedimentary succession along the eastern margin of the basin, and displaced basement by several kilometres (Kamp, 1991). The Patea-Tongaporutu High was accentuated as a basement ridge at this time. Geohistory plots show that by the Early to Middle Miocene, much of the study area was residing at mid to lower bathyal depths (Fig. 7.6a,b). In fact, water depths adjacent to the Taranaki Fault were greater than at any other time in the basin's history.

The Late Oligocene-Early Miocene change from carbonate to terrigenous sedimentation in Taranaki Basin marked the beginning of a major 1st-order regressive megacycle, resulting in a northwestward progradation of the shelf wedge that is still continuing today. By the Late Waiauan-Early Tongaporutuan (12-10 Ma), the Taranaki Fault was less active (King and Thrasher, 1996), and sediment supply began to outpaced subsidence rates along Taranaki Basin's southern and southeastern margins. The paleoshelf at this stage was inland (east) of the modern-day shoreline; studies of Tongaporutuan-aged Mount Messenger Formation sediments cropping out in cliffs along the modern-day coast (e.g., King et al., 1993, 1994; Hansen, 1996) indicate that the shelf break must have been to the east and south of these basin floor fan and slope deposits. Analysis of Wanganui and King Country Basins has established the Late Miocene shoreline and approximate shelf edge position in these basins (Kamp et al., 2002).

Kamp et al. (2002) identified four 2nd-order regressive cycles in King Country and Wanganui Basins superimposed on the Miocene regressive megacycle. The Mount Messenger, Kiore (Vonk et al., 2002), and Urenui Formations accumulated in a slope to basinal continental margin (King and Thrasher, 1996; Vonk et al., 2002), and together with the shelfal Matemateaonga Formation, constitute the third of these four 2nd-order cycles (Whangamomona Group of Kamp et al., 2002; Vonk et al., 2002). This Group was extensive throughout Wanganui and King Country Basins, and progressively overtopped the Patea-Tongaporutu High, encroaching into the eastern margin of Taranaki Basin, but had only a limited extent into western parts of Taranaki Basin (Fig. 7.7).

A main feature of northern Taranaki Basin geography through the Late Miocene was a series of submarine stratovolcanoes, forming a semi-linear north-northwest trend along the axis of what was to become the Northern Graben. The main period of active volcanism was between 14 and 11 Ma (King and Thrasher, 1996); episodes of Late Miocene volcanism are evidenced by the occurrence of sporadically to highly volcaniclastic sediments inter-fingering with non-volcaniclastic mudstone, siltstone, and minor sandstone observed in almost all well sections throughout northern Taranaki Basin. Volcaniclastic sediment formed 'aprons' around the base of volcanic massifs. Deposition into deeper waters probably occurred in the form of turbidity currents and by other mass flow mechanisms, with some subaqueous settling through the water column. The gradual transition to the dominance of terrigenous mud and less frequent occurrence of volcaniclastic beds upwards in the sediment pile points to a waning of volcanism in northern Taranaki Basin.

Sediment deposition in northern Taranaki Basin during the Tongaporutuan and into the Kapitean was clearly influenced by proximity to these volcanic edifices. Well sections located closer to volcanic centres have a stratigraphy dominated by volcaniclastic sediment (Mohakatino Formation; e.g., Ariki-1, Tangaroa-1), while those further away from active centres are dominated by hemipelagic mudstone (Manganui Formation; e.g., Taimana-1, Arawa-1).



7.6: **(a)** Earliest Miocene and **(b)** Middle Miocene paleogeographic map (redrawn and modified from King and Thrasher, 1996).

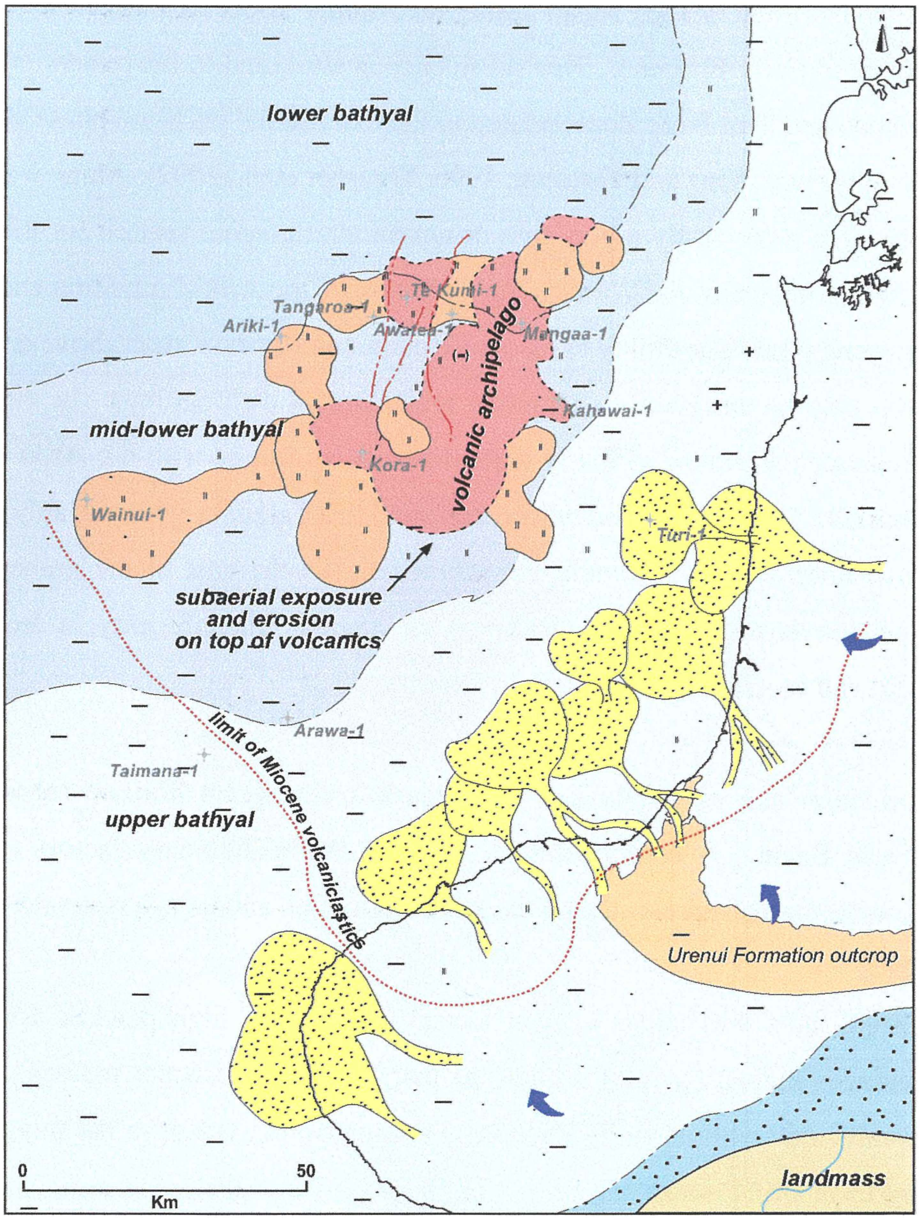


Fig. 7.7: Inferred Late Tongaporutuan to Early Kapitean (c.8 Ma) paleogeography. Volcaniclastic sediment derived from volcanic massifs and redeposited down flank, often over large distances. Deposition of terrigenous sediment in fans dominantly confined to a zone paralleling the modern day shoreline. Entire study area under bathyal water depths and marginal oceanic to oceanic water mass conditions.

Several of these volcanic massifs formed topographic highs into the Pliocene, resulting in localised unconformities between the Miocene volcanics and overlying Pliocene sedimentary section. Isopach maps illustrate that volcanic edifices often continued to exert an influence on the distribution and thickness of sediment well into the Pleistocene.

7.4.3 Late Kapitean to Early Opoitian (Fig. 7.8: c.6 – c.4 Ma)

Subsidence of the Northern Graben began during the Middle Miocene, but accelerated during the latest Miocene to Early Pliocene. This subsidence is attributed to movement on the Cape Egmont Fault Zone and Turi Fault Zone, related to the extensional back-arc phase of Taranaki Basin's structural history (King and Thrasher, 1996; Thrasher et al., 2002). Many well sections within the study area record only a very thin or absent stratigraphic section for this interval. Over the northwestern parts of the Western Stable Platform, the Ariki Formation encompasses much of this missing time. The ability to recognise certain intervals of time above or below the Ariki Formation may be attributed to sampling resolution of well cuttings. In the Northern Graben, two separate calcareous to marly units that overlap in age with the Ariki Formation have been identified. These condensed units correspond to a Taranaki-wide variably developed hiatus between Miocene and Pliocene-aged sediments. To the east of the graben, missing stratigraphic section is expressed physically as an angular unconformity, a result of the significant uplift and erosion as part of graben formation.

The Ariki Formation and age-equivalent units mark a significant horizon across parts of northern Taranaki Basin. A combination of tectonic and sedimentary factors combine to explain the paraconformity represented by the Ariki Formation and its age equivalents with the Northern Graben:

- The study area was under bathyal water depths, with high planktic foraminiferal percentages indicating fully oceanic to marginal oceanic water masses. This may have coincided with an Early Pliocene warming trend evident in the oxygen isotope record from DSDP Site 593.
- While most volcanism had ceased or migrated to the south by the Kapitean, parts of major volcanic massifs were still emergent (above seabed), determining to some extent sediment pathways, and possibly providing an effective barrier to westward sediment transport.
- The total subsidence curve for Ariki-1, Tangaroa-1, and Wainui-1, and palinspastic restoration of seismic reflection profile P95-158, indicate that near the end of the Miocene, some uplift did occur on the northwestern part of the Western Stable Platform, forming a low-relief paleohigh. This paleohigh probably reflects down-faulting of the Northern Graben, with uplift to the east of the graben even more pronounced (Fig. 7.5). Geohistory plots reveal that Mangaa-1 and Awatea-1, located in the graben, obtained their greatest water depths by the Early Pliocene.

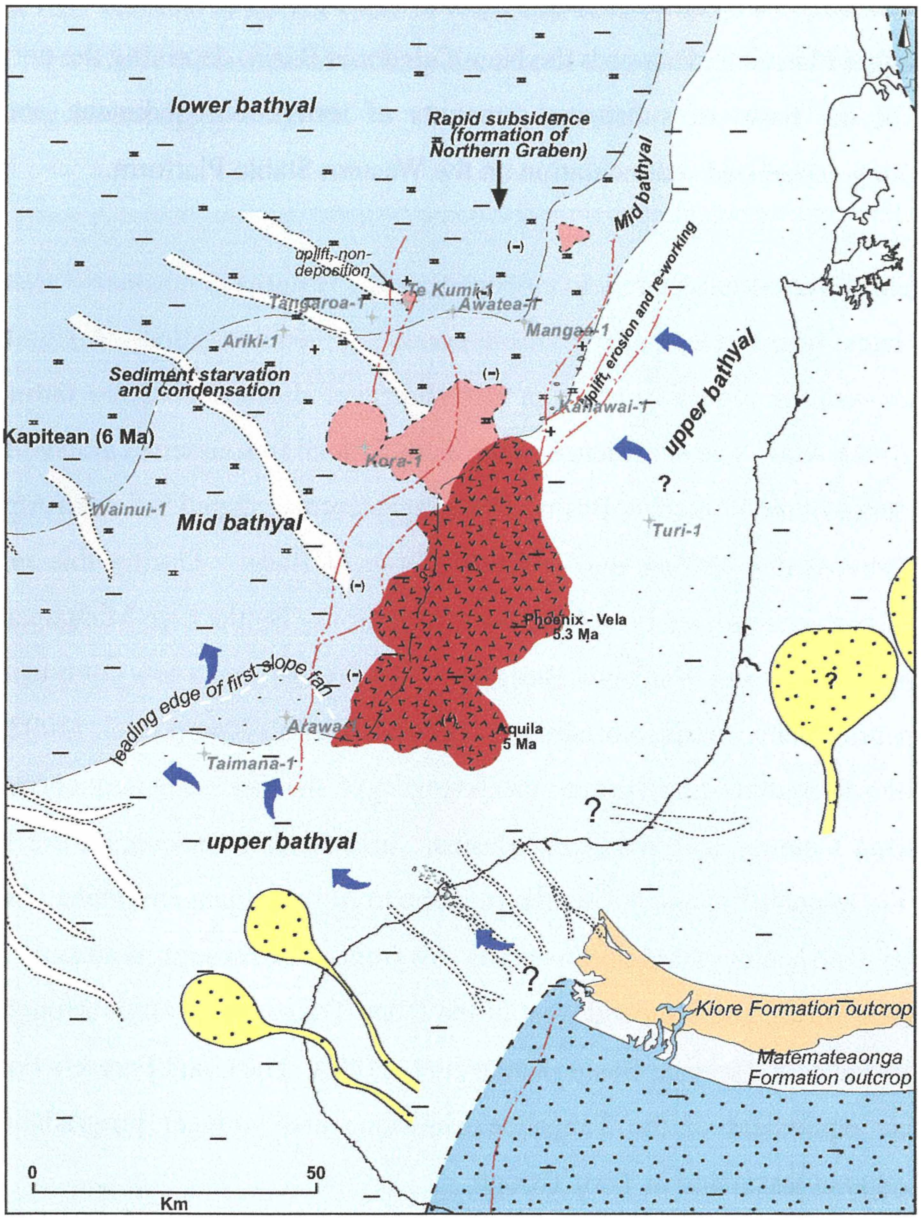


Fig. 7.8: Inferred Late Kapitean to Early Opoitian (c.5 Ma) paleogeography. General sediment starvation throughout the northern part of the study area, leading to condensed and marly Ariki Formation and equivalents. Some volcanism from Aquila and Phoenix volcanic centres. At this time water depths were still bathyal, but beginning to shallow to the south.

- Uplift along Kahawai Fault during the latest Kapitean to Early Opoitian, and the derivation of sediment from the east, resulted in the deposition of Mangaa Formation sandstone beds in the Northern Graben. During an interval of interrupted sediment supply, a marly unit equivalent in age to the upper part of the Ariki Formation accumulated in the graben.

- Large channels systems (top Miocene channels mapped by Thrasher and Cahill, 1990; Fig. 2.5) may have directed South Island-derived sediment northward across the Challenger Plateau and towards the New Caledonia Basin, depriving the northwestern areas of the basin of substantial amounts of terrigenous sediment, and thereby promoting condensed sedimentation on the Western Stable Platform.

The factors listed above pertain mainly to paleogeographic features in northern Taranaki Basin that promoted latest Miocene to Early Pliocene condensed sedimentation. A significant and additional factor was the rapid formation of the Wanganui Basin depocentre during the late Early Pliocene (~4.8 Ma). The subsidence of the Toru Trough in southern Taranaki Basin was linked to subsidence of the Wanganui Basin. These depocentres trapped Southern Alps-derived sediment and prevented it getting into northern Taranaki Basin. During this interval the Tangahoe Mudstone accumulated in bathyal water conditions. By the Late Mangapanian, shelf conditions were established in Wanganui Basin and Toru Trough, and a new continental margin succession was prograding north into northern Taranaki Basin. Kamp et al. (2002) grouped these units in the Rangitikei Supergroup, the youngest of the four megasequences that they identified in King Country and Wanganui Basins. Immediately underneath the Rangitikei Supergroup is the Matemateaonga Formation (upper part of the Whangamomona Group). The Rangitikei megasequence migrated northward as two fronts, one through Wanganui Basin and into King Country Basin, and a second west of the Patea-Tongaporutu High through the Toru Trough and into northern Taranaki Basin (Kamp et al., 2002). The Giant Foresets Formation is the stratigraphic equivalent of the Tangahoe Mudstone and younger progradational shelf deposits in Wanganui Basin and in Toru Trough.

7.4.3 Late Opoitian (Fig. 7.9: c.4 - c.3.5 Ma)

The mid to Late Opoitian was characterised by continued sediment starvation on the northwestern Western Stable Platform. In contrast, to the south, in the vicinity of Arawa-1 and Taimana-1, sediment accumulated as a series of large lobate slope fans evident on seismic reflection profiles. These slope fans marked the beginning of progradation of the Giant Foresets Formation sedimentary wedge into northern Taranaki Basin. In addition to these fans, the newly created accommodation in the Northern Graben attracted an influx of coarse (sandy) terrigenous sediment, well expressed on the wireline logs obtained for Awatea-1 and Mangaa-1 (Mangaa Formation). This sediment was probably sourced mainly from the east. Erosion of

Whangamomona Group strata after the Late Opoitian is also considered to be a contributing source for Late Pliocene to Recent strata in Wanganui Basin (Kamp et al., in prep.). A direct Southern Alps source is considered unlikely because of the sediment trap created in Wanganui Basin during the late-Early Pliocene.

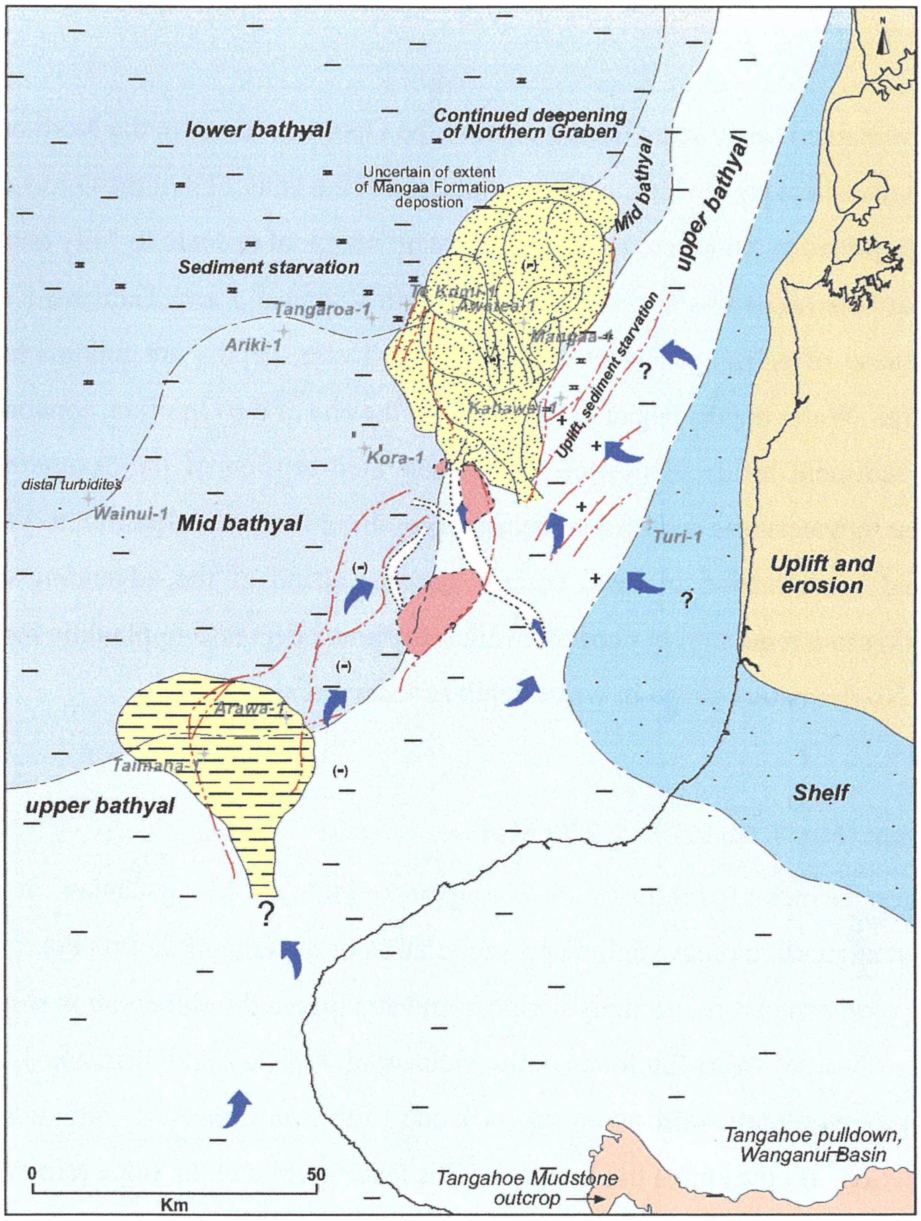


Fig. 7.9: Inferred Late Opoitian (c.4 Ma) paleogeography. Formation of the Northern Graben created accommodation in which coarse terrigenous sediment accumulated (Mangaa Formation). To the south, slope fans prograded northward into the study area, while the northwestern region, particularly in the vicinity of Tangaroa-1 and Te Kumi-1, was still subject to sediment starvation (Ariki Formation). At this time, bathyal depths still prevailed.

While sediment was accumulating fairly rapidly in the Northern Graben (>0.5 m/1000 yrs.), other sites in the northern part of the study area were still starved of terrigenous sediment.

Wainui-1 was dominated by hemipelagic muds, but was beginning to receive slightly coarser material, probably via distal turbidites initiated ahead of the advancing slope front. Ariki-1 also received a limited amount of mud during the Late Opoitian, although slightly further to the north and east, both Tangaroa-1 and Te Kumi-1 were still subject to sediment starvation. Terrigenous sedimentation associated with the Mangaa Formation is abruptly terminated to the west by Te Kumi Fault.

Water depths remained fairly static during most of the Opoitian. Even in the Northern Graben, sedimentation rates kept pace with subsidence throughout the latter half of the Opoitian. Depths are mainly interpreted as middle to lower bathyal with marginal oceanic to fully oceanic water masses. Upper bathyal depths are interpreted for both Kahawai-1 and Taimana-1 during this interval. Because of subsequent erosion at the Turi-1 site, depths are unknown, but were probably shelfal. Water depths began to shallow near the end of the Opoitian, concomitant with an increased sediment influx associated with accelerated erosion of the Southern Alps. A marked change in watermass conditions (oceanicity) during the latest Opoitian to Waipipian is possibly linked to deflection of local current systems ahead of the advancing continental margin, resulting in a reduction in nutrient availability and a decrease in planktic foraminiferal percentages. No dramatic change in water depth is observed at this time.

7.4.4 Waipipian (Fig. 7.10: c.3.5 - c.2.79 Ma)

Sedimentation rates increased through the Waipipian and into the Mangapanian. It was during this period that aggressive progradation and aggradation of the Giant Foresets Formation really began. In the southern part of the study region, rapid and directed sedimentation resulted in the continued deposition of slope fan lobes in the vicinity of Arawa-1 and Taimana-1. Fan lobes continued to prograde northward, swinging back and forth from a westerly direction to a more northerly direction. By the end of the Waipipian, the leading edge of the slope fans was between Taimana-1 and Wainui-1, with the apex of the fans migrating from south of Taimana-1 to northwest of this site.

Terrigenous deposition was now occurring across the study area, although in the northern region (vicinity of Tangaroa-1 and Te Kumi-1), only a very thin Waipipian succession (early bottomsets of the Giant Foresets Formation) is recorded. Slightly thicker Waipipian sequences accumulated over the northwestern parts of the Western Stable Platform (e.g., Ariki-1, Wainui-

1). Mixing of inner to mid shelf benthic foraminifera with deeper water species at Ariki-1 and Wainui-1 demonstrates that there was active mobilisation and transport of sediment from shallow to deeper water environments.

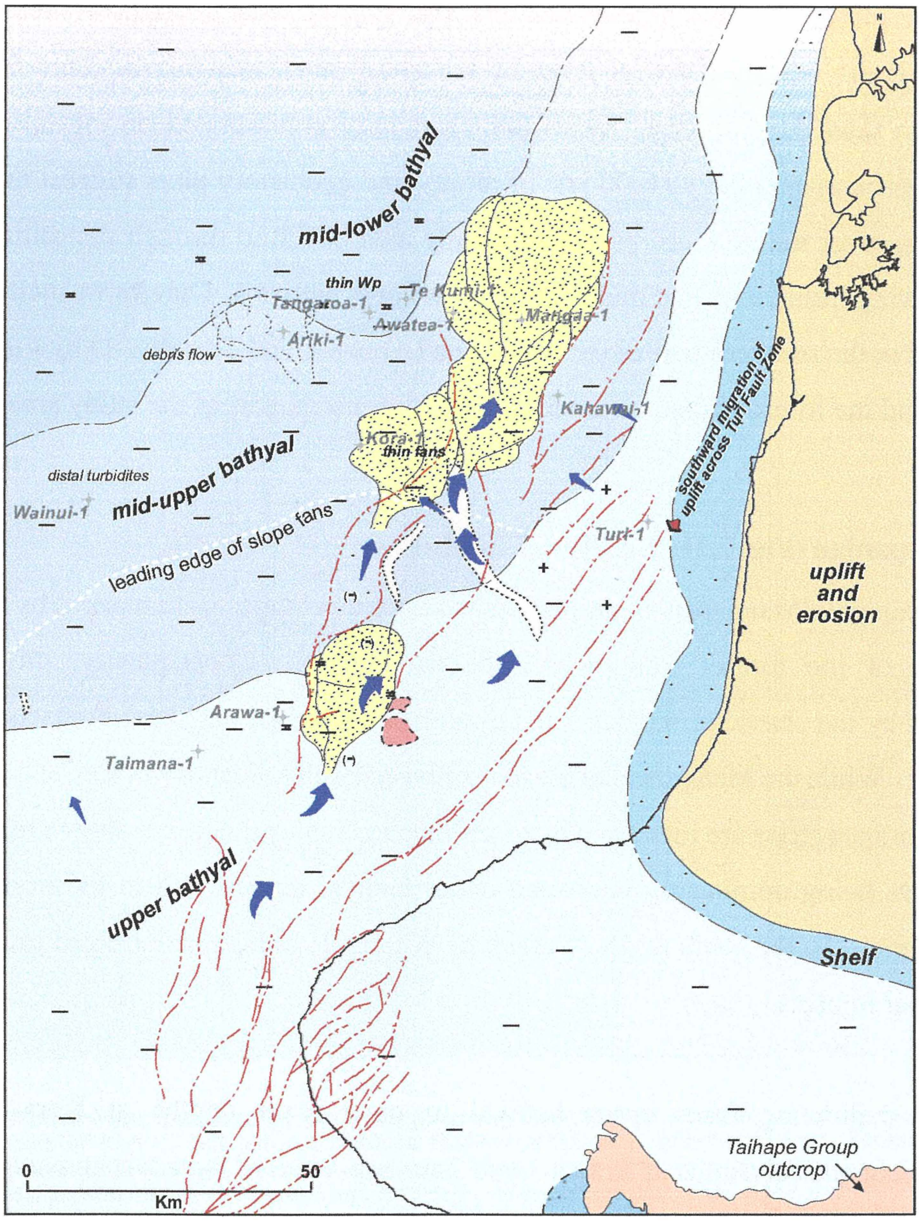


Fig. 7.10: Inferred Waipipian (c.3 Ma) paleogeography. The contemporary shelf break had prograded past Taimana-1 and Arawa-1 by this time. Sediment deposition is focused between Taimana-1, Arawa-1, and Wainui-1, and in the Northern Graben. Water depths begin to shallow rapidly.

Thickest accumulations of Waipipian strata occur within the Northern Graben, particularly to the east of Arawa-1, and in the southwestern quadrant of the Western Stable Platform. Sandstone beds were still well developed during much of this interval in the vicinity of Mangaa-1 and Awatea-1, although towards the end of this stage (particularly evident in the Awatea-1 well section) became muddier and less well developed. This suggests that the feeder system supplying sediment to these fan lobes was beginning to be shut off.

By the end of the Waipipian Stage, the additional accommodation created in the Northern Graben was rapidly being filled, and sediment was spilling over to adjacent areas, although bathyal conditions persisted across most of the study area west of the Turi Fault Zone. Sedimentation rates for bottomset facies (which mainly accumulated during this stage) are moderate (approximately 0.5 m/1000 yrs. or more), and geohistory plots suggest that much of the shallowing over the northern part of the study area occurred through deposition of these units (shallowing from lower and middle bathyal to upper bathyal). Outermost shelf conditions were reached in the southern part of the study area (Arawa-1 and Taimana-1) by the end of the Waipipian, and the foreset front had advanced into the central part of the study area.

7.4.6 Mangapanian (Fig. 7.11: c.2.79 - c.2.28 Ma)

The latest Waipipian/Mangapanian through to Nukumaruan Stages were marked by accelerated progradation of the foreset front across the study area. Consequently, this period is characterised by the classic progradational clinoforms that make the Giant Foresets Formation so distinctive. While the Mangapanian Stage is often difficult to identify in well sections, where Mangapanian-aged strata are identified they continue to display the trends shown by Waipipian strata, such as being dominantly deposited under bathyal depths, and in the north, forming bottomsets units. To the south (in the vicinity of Wainui-1) Mangapanian-aged strata occur as progradational foresets.

Progressive shallowing from upper bathyal to outer shelf depths at Kora-1, and the establishment of shelfal depths at Arawa-1 and Taimana-1 record the advancement of the shelf margin past these sites. The shelf margin similarly advanced past Kahawai-1 on the fringe of the Turi Fault Zone, indicating that progradation was advancing on two fronts – one from the south and one from the east. This suggests that sediment sourced from the more proximal North Island was contributing to progradation.

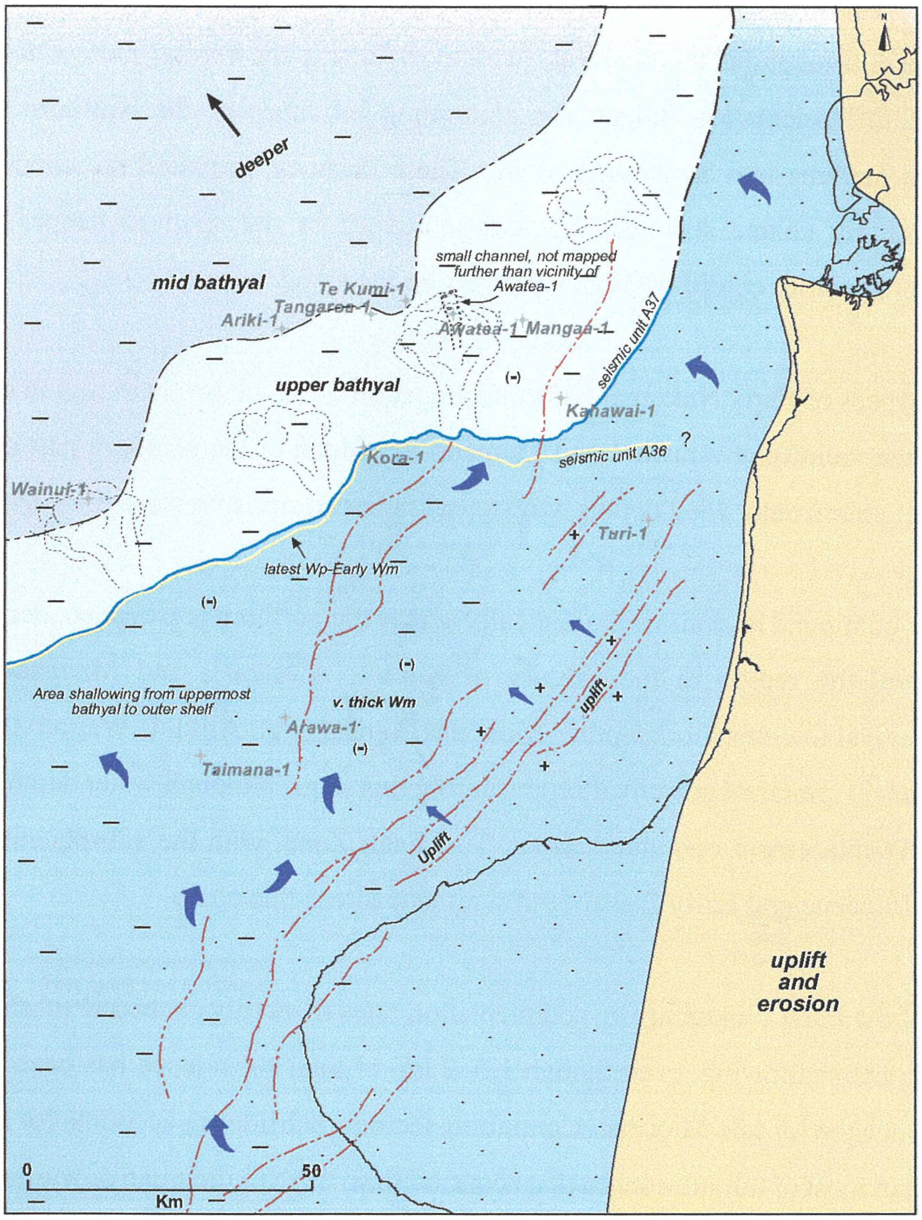


Fig. 7.11: Inferred Mangapanian (c.2.5 Ma) paleogeography. Most of the sediment is tied up in a slope front south of Kora-1, although the Northern Graben continues to attract sediment. Turbidity currents generated downslope from the advancing foreset front contribute to thicker accumulations of Mangapanian strata at Wainui-1 and Awatea-1. The southern part of the study area shallowed rapidly from uppermost bathyal to outer/mid shelf following progradation of the shelf margin into northern regions.

7.4.7 Early Nukumaruan (Fig. 7.12: c.2.28 - c.1.8 Ma)

The Early Nukumaruan was a period of significant sedimentation throughout the study area, recording the rapid progradation and aggradation of the Giant Foresets Formation. Through this period the continental margin advanced rapidly, and by this stage the leading edge of the foreset front had reached the northernmost region of the study area. Isopach mapping has revealed a thick linear-shaped lobe of Early Nukumaruan sediment that extends from the west (Wainui-1) to the northeast (north of Mangaa-1). Numerous channel networks have also been mapped over the seismic grid throughout this interval, particularly along the central axis of the Northern Graben. Channel orientation reflects the continuing influence of the Northern Graben on sedimentation patterns into the Nukumaruan. These channels provided the conduit through which increasingly coarse sediment was transported from the shelf to deeper areas, and deposited in channel-levee and overbank deposits on the slope.

Thinner sequences of Early Nukumaruan strata are recorded in the southern region of the study area (i.e., in the vicinity of Arawa-1 and Taimana-1). Much of the southern part of the study area resided at mid to outer shelf depths, with topset facies comprising the sedimentary section.

Water depths continued to shallow dramatically across the northern region also, and by the end of this interval the region in the vicinity of Kora-1, Awatea-1, and Mangaa-1 obtained uppermost bathyal to outer shelf depths. More distal areas (e.g., Ariki-1, Wainui-1, Tangaroa-1 and Te Kumi-1) remained at bathyal depths during this stage, although water depths continued to shallow. Displacement continued on the Turi Fault Zone, with the subsequent erosion of much of the Pliocene and Early Pleistocene sediment across this zone.

By the end of the Early Nukumaruan, sedimentation rates markedly exceeded subsidence rates, resulting in a net shallowing, even though 1.5–2 km of total subsidence has been recorded at many sites since the Middle Miocene. Continuing tectonic subsidence accounts for some of this subsidence, but most of the subsidence that occurred from the Nukumaruan onwards in the north (earlier in the south) occurred as a result of sediment loading as the progradational wedge (Giant Foresets Formation) moved through the study region. Geohistory plots illustrate that sites shallowed and total subsidence increased as sediment began to accumulate more rapidly, particularly as the continental margin advanced past a site. After the foreset front passed each site, and the site shallowed to shelf depths, deposition occurred at slower rates (topset strata), with sediment accumulation approximating subsidence rates.

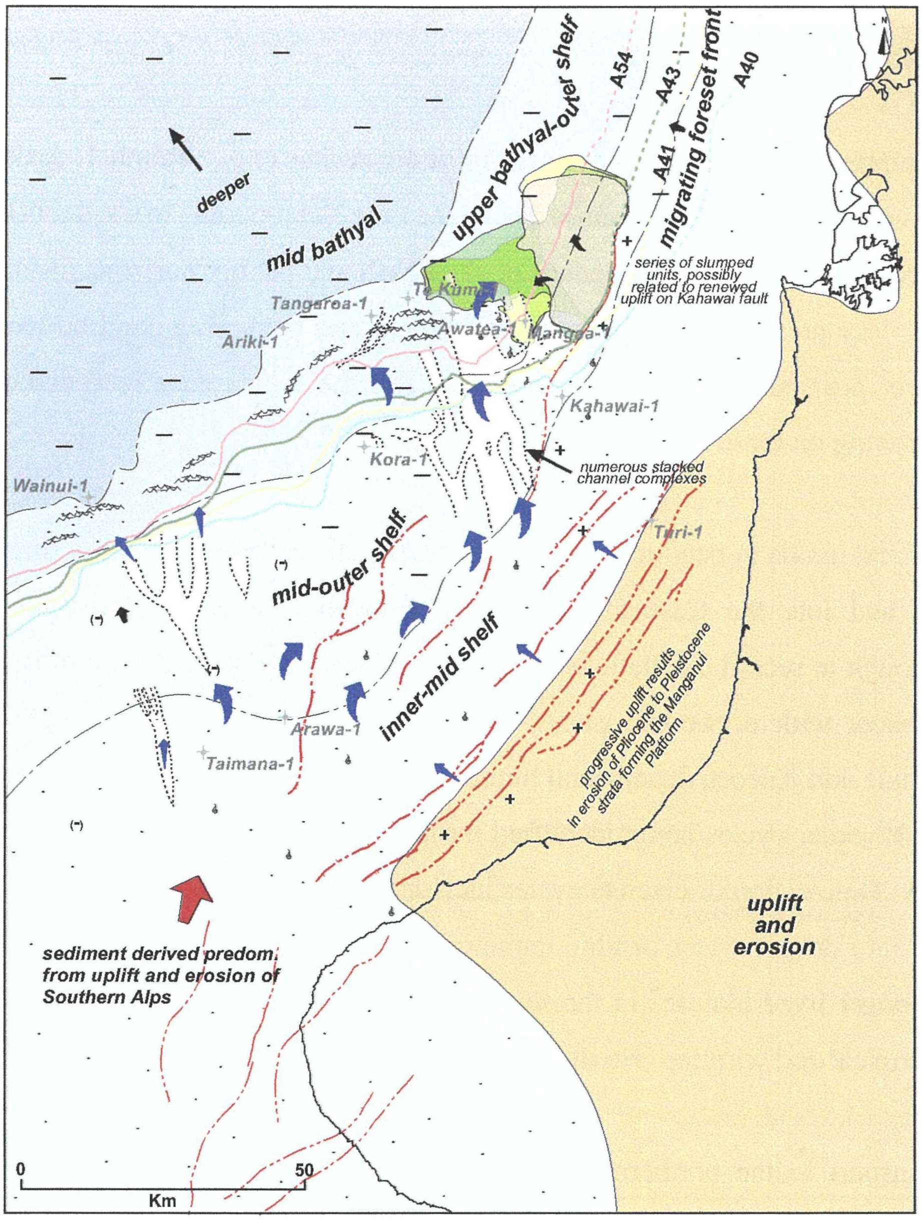


Fig. 7.12: Inferred Early Nukumaruan (c.2 Ma) paleogeography. The progradational wedge rapidly advanced across the study area, and water depths shallowed from predominantly bathyal to predominantly uppermost bathyal to outer shelf in the north. Channel systems delivered sediment to much of the northern and northwestern region of the study area. Shelf breaks of seismic units A40 (oldest), A41, A43, and A54 (youngest) are show to illustrate prograding continental margin. The shelf break at this time is beginning to approximate the position of the modern day shelf break.

7.4.8 Late Nukumaruan to Castlecliffian (Fig. 7.13: c.1.8 - c.0.33 Ma)

The shelf continued to broaden throughout the Late Nukumaruan and into the Castlecliffian, being a direct response to the flux of sediment derived from uplift and erosion of the Southern Alps (Kamp et al., 1989). The progradational wedge advanced rapidly northwards and westwards across the study area, reaching the northwestern-most regions by the end of the Nukumaruan. The thickest part of the Late Nukumaruan-Castlecliffian succession is associated with this wedge.

Benthic foraminiferal associations at even the most distal sites (e.g., Wainui-1, Ariki-1) record an increasing content of inner shelf faunas during the Late Nukumaruan to Castlecliffian. Many sites also record an increase in the amount of shell hash and pebbly horizons identified in the well sections. The prevalence of shallow water faunas and increasing contributions of shelly material reflect proximity to the paleo-shoreline, and the increased effect shallower water depths were having on these more distal sites.

Water depths and trends in benthic assemblages display much more variability throughout the Nukumaruan and into the Castlecliffian than previously, and suggest that water depths shallowed enough to record oscillations in global sea level. Marginal neritic to neritic water-masses prevailed, with occasional pulses of marginal oceanic conditions. By combining seismic, wireline and paleoenvironmental trends, a degree of cyclicity can be determined, with a number of 4th-order cycles being identified through the Nukumaruan and lower part of the Castlecliffian. Delineation of discrete cycles includes the identification of pulses of shallower water faunas into deeper water benthic communities, lower planktic ratios, identification of diastems or longer lived hiatuses in the seismic record, as well as other seismic criteria, and various lithological and wireline criteria.

Sediment transport to the northern Taranaki Basin during the Pleistocene may have been seriously compromised by coastal dynamics during high sea levels. During glacial minima, sediment is modelled as being transported around a developing Farewell Spit or equivalent, and into the early Cook Strait, much as it is today. However, during lower sea levels the emergence of a land bridge between the North and South Islands, connecting Taranaki Peninsula with Cape Farewell, altered tidal flow and sediment distribution patterns, diverting sediment to the west and north (Proctor and Carter, 1989). This scenario is consistent with the implication that much of the outbuilding and widening of the continental shelf occurred during Pleistocene glaciations, with an increased sediment flux derived from uplift and glacial erosion carried northwards by the D'Urville Current.

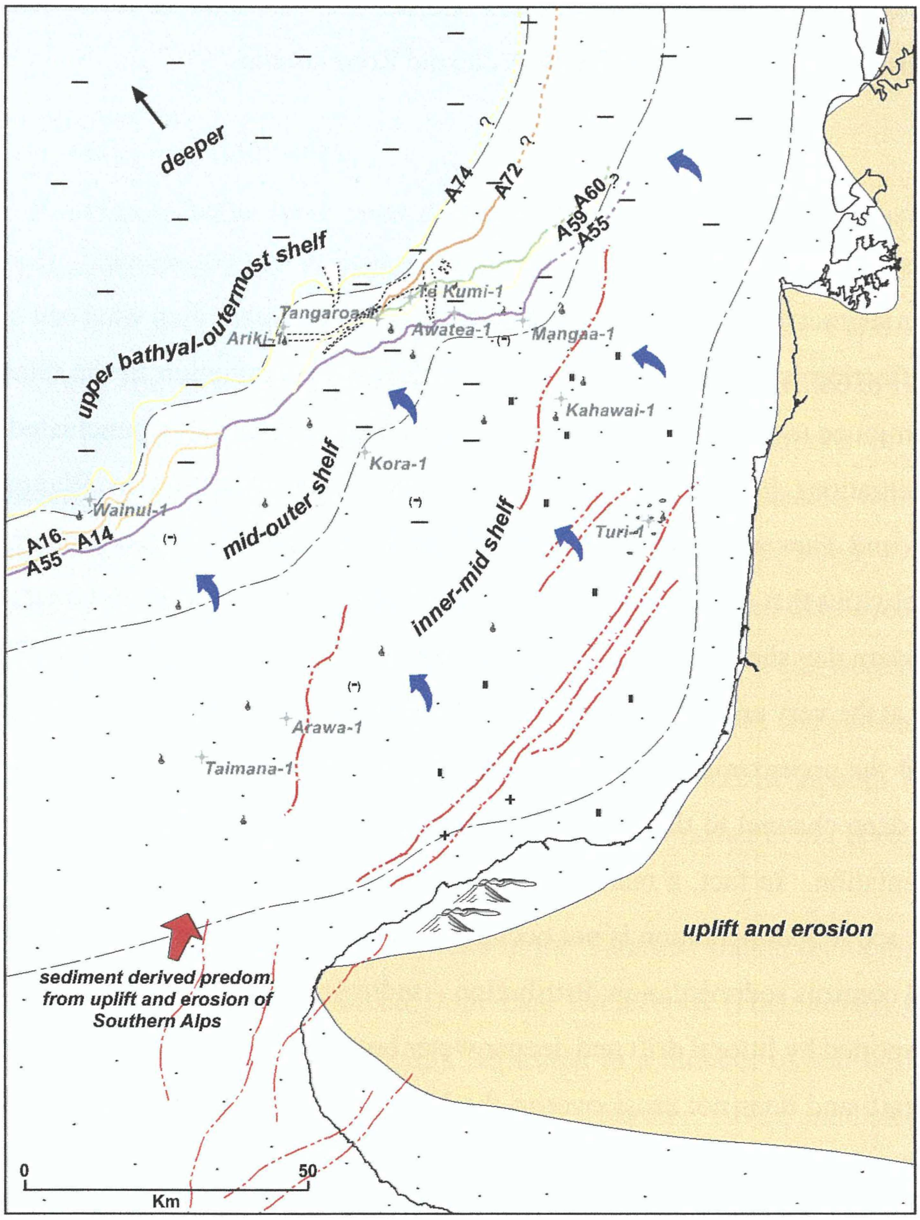


Fig. 7.13: Inferred Late Nukumaruan to Castlecliffian (c.1 Ma) paleogeography. A Late Nukumaruan phase of retrogradation results in a short-term landward movement of the shelf break, before progradation (degradational foresets) continues to the north and west. Progradation rates dropped off in the Castlecliffian, possibly due to capture of sediment by Cook Strait. Dominantly middle to outer shelf depths, with marginal neritic to neritic water masses. Shelf breaks of seismic units A55 (oldest), B14, B16, A59, A60, A72, and A74 (youngest) are shown to illustrate prograding continental margin.

The latest foreset units to be deposited have a distinctly different seismic character from underlying foreset units. These younger units become increasingly degradational, incising deeply into older (mainly lower Nukumaruan) strata. Well completion reports indicate a physical change in the properties of the mud between the older progradational foresets, and the younger degradational foresets, as well as an increasing shell hash component. While no detailed provenance studies have been undertaken, it is suggested that these characteristics reflect an increasing influence from sediment sourced from erosion of North Island's axial ranges, and possibly a volcaniclastic Taupo Volcanic Zone source.

7.4.9 Recent

At most well sites, the first 200-400 m of sedimentary section was not sampled. Hence it is not possible to reconstruct the period from Castlecliffian to Recent, other than what can be observed on seismic reflection profiles. Water depths of sediment accumulation in the Giant Foresets Formation continued to shallow to the present day bathymetry, invariably punctuated by regular sea level fluctuations linked to global climate change (as identified in Wanganui Basin Castlecliffian and Haweran strata; e.g., Naish and Kamp, 1997). From seismic reflection profiles, it is obvious that the focus of sedimentation had now moved north and west of the study area. The modern day shelf break can be seen on seismic reflection profiles P95-158, P95-168, P95-103, and at the very end of AR89-446-108. Modern-day bathymetric maps show a number of incisions at the approximate present day shelf break; one of these incisions, which can be traced into a deep channel to the west of Tangaroa-1 on line P95-158, shows no evidence of active sedimentation. In fact, a marked seismic reflection couplet at the modern-day seabed indicates that active sedimentation is not occurring in this area at present. The present relative high sea level controls sedimentation distribution – sediment derived from erosion of the South Island is transported by littoral drift and deeper water bottom currents around Farewell Spit and into Cook Strait, and does not cross over to the North Island (Williams, 1985). Much of the sediment derived from erosion of the North Island ranges is also captured in littoral drift systems, and directed either north or south of the Taranaki Peninsula, although some sediment may reach the outer margin of the shelf via deeper water circulation patterns.

7.5 Petroleum systems of northern Taranaki Basin

The Giant Foresets Formation has not been the target of active hydrocarbon exploration in the Taranaki Basin, nor have any significant oil shows ever been recorded within this formation. However, because of the great thickness and rapid deposition of the formation, there has been considerable impact on underlying petroleum systems, particularly on the thermal regime, and there is also the suggestion that certain structures within the foresets may have the possibility of being charged within the last few million years. This section summarises the work of various authors whom, over the past several years, have investigated this aspect of the Giant Foresets

Formation. For a complete summary of Taranaki Basin petroleum systems, refer to King and Thrasher (1996).

7.5.1 Sources, seals, and reservoirs

More than 20 offshore wells have been drilled in the northern Taranaki Basin since 1970 (Mangaa-1), sixteen in the 1980's (Bennett et al., 1992). All of these wells have been plugged and abandoned, most as dry holes. Oil shows were however noted in Turi-1 and Tangaroa-1, as well as Awakino-1 and Mokau-1 to the east of the study area. Kora-1, drilled in 1987 by ARCO Petroleum NZ Inc., contained significant sub-commercial accumulations of hydrocarbons within pyroclastic and epiclastic rocks (Mohakatino Formation) associated with the Miocene Kora volcanic structure (testing at 1168 BOPD; Bergman et al., 1992; Reed, 1992). A lower target, Eocene-aged Tangaroa sands within the Tangaroa Formation, displayed hydrocarbon shows but yielded nothing substantial on testing. A number of other wells were drilled into the Kora volcanic structure (Kora-2, 3, and 4), but no significant oil shows were encountered in any of them. However, exploration interest has recently been rekindled in both this and other volcanic structures in the northern Taranaki Basin (e.g., see Thrasher et al., 2002).

Hydrocarbons extracted from commercial wells in the Taranaki Basin, including the Western Stable Platform, are most likely sourced from Late Cretaceous (Haumurian) terrestrial and shallow marine sediments (Pakawau Group), deposited in a number of sub-basins associated with opening of the Tasman Sea (Holt and Stern, 1991; Thrasher, 1992), and Paleogene Kapuni Group sediments (Reed, 1992). Geochemical characteristics of upper and lower Late Cretaceous coal measures of the Pakawau Group have total organic carbon (TOC) averages of 10% and 9%, respectively. Biomarkers indicate that the coals were primarily derived from gymnosperm (conifer) type organic matter. In comparison, upper Late Cretaceous shallow marine rocks have a TOC average of only 0.4% (where 1% is considered the cut off point for rocks with possible source potential). Biomarkers from oils generated from the Kapuni Group indicate that they are derived from angiosperm (flowering plant) types of organic matter (Reed, 1992).

Late Cretaceous rocks were encountered at both Ariki-1 and Wainui-1; seismic mapping by Thrasher (1992) has shown equivalent aged coal measures to be extensive over the Western Stable Platform, forming a veneer up to 50 m thick. Other areas of the northern basin however are thought to be underlain by a wholly marine sequence. While Late Cretaceous source rocks are inferred (Reed, 1992), major distinctions between the hydrocarbons encountered in commercial wells to the south, and those from Tangaroa-1 and Kora-1 are recognised (see King and Thrasher, 1996, their Figs. 6.14 and 6.15, pgs 138 and 139 respectively). Analyses of hydrocarbons from Tangaroa-1 and Kora-1 reflect a significant marine influence (Johnston, 1991, in Sykes et al., 1992) compared with hydrocarbons generated from either the Pakawau or Kapuni group coals. Large amounts of marine algae biomarkers (Reed, 1992), which have a heavier carbon isotopic composition, as well as a moderate amount of oleanane (a biomarker associated with angiosperms), are present. These indicators suggest that the oil encountered in both Tangaroa-1 and Kora-1 represent generation from the same source (Reed, 1992). The presence of oleanane and heavier carbon isotopic composition indicates some influence from terrestrially derived material, and a very shallow (nearshore or estuarine) depositional environment with limited circulation (Reed, 1992), consistent with a shallow marine environment as interpreted by Thrasher (1992). Though not definitively identified, probable source rocks are Late Cretaceous coal measures and Paleocene shales (Thrasher, 2002). Johnston (1992, in Bennett et al., 1992) concluded that the source for Awakino-1 hydrocarbons is again different from the Kora or Tangaroa source, suggesting that these two areas are underlain by different source rocks, or do not have common migration pathways.

To date, all of the commercial hydrocarbon accumulations in Taranaki Basin are contained within Neogene structural traps within the Eastern Mobile Belt (Bennett et al., 1992; King and Thrasher, 1996). While the Kora hydrocarbon accumulations are in volcaniclastic sediment onlapping the flanks of the Kora volcanic complex, charge of this reservoir is considered to have been intimately linked to Neogene movement on faults associated with the Cape Egmont Fault Zone (Reed, 1992). However, this petroleum reservoir is very complex, and is considered to have unpredictable production behaviour due to heterogeneous stratigraphic relations, limited lateral continuities of units, and complicated diagenesis and pore character (Bergman et al., 1992). Rather than reservoirs associated with volcanic structures, Bergman et al. (1992) and Reed (1992) consider the most prospective reservoirs to be deeply buried volcaniclastic sequences of the Mohakatino Formation and equivalents within the axis of the Northern Graben, which occur over some 25% of the graben area.

Thrasher et al. (2002) recognise two distinct categories of traps for prospects identified within the Northern Graben, including those associated with graben forming tectonics, and those associated with igneous centres. Potential reservoir units include Eocene channel and basin floor complexes of the Mangahewa, McKee and Tangaroa formations, the volcaniclastic Mohakatino Formation, and submarine fans sands of the Moki (Early Miocene), Mount Messenger (Mid-Late Miocene) and Mangaa (Pliocene) Formations (King, 1992; Thrasher et al., 2002). King (1992) suggests that lateral facies changes in the offshore region may limit the Moki and Mount Messenger Formations as potential reservoirs; Mount Messenger Formation sediments are only encountered as far west as Turi-1 and thus do not constitute a serious play in the study area. Well-developed age equivalent sands encountered at Awatea-1 and Mangaa-1 (C1 sands of Forder and Sissons, 1992, intra-Manganui sandstone and Mohakatino Formation, respectively in this study) are considered to be a major play in the Northern Graben. They are mappable over a large area, display a blocky wireline motif characteristic of sand-dominated basin floor fan deposition, and are overlain by a thick fine-grained seal. Younger Pliocene-aged Mangaa A and B sands (C2 and C3 of Forder and Sissons, 1992) also display a wireline motif characteristic of basin floor fan deposition, and, like the older sands, form an elongated clastic wedge oriented parallel to the axis of the Northern Graben, with its depositional extent constrained to the east and west by major faults (see Fig. 5.7a,b).

Hydrocarbons have been found in Latest Miocene to Early Pliocene shelf sandstone beds of the Matemateaonga Formation (King and Thrasher, 1996), and King (1992) suggests that younger (Plio-Pleistocene) deep-water analogues of the Mount Messenger and Moki sands may be highly prospective. In the study area, the probability of hydrocarbons accumulating in similar shelf to basin floor sandstone beds of the Giant Foresets Formation is variable, depending on maturation and timing of expulsion of hydrocarbons (discussed in following section). To date, no hydrocarbons have been found in any structures associated with deposition of the foresets or Plio-Pleistocene fault movement, and the Giant Foresets Formations themselves are more commonly credited for acting as a regional seal rather than reservoir. However, seismic reflection profiles provide indications that there is some migration up through the formation, evidenced by numerous 'bright spots' on the profiles (for example, immediately to the west of Kahawai-1, at a depth of ~750 msecs, Enclosure 3). Thus there is potential for hydrocarbon entrapment, given a suitable reservoir and appropriate timing of migration. There are however some major considerations that need to be properly evaluated when assessing the hydrocarbon prospectivity of the Giant Foresets Formation.

- Oil expulsion and migration to the Kora structure is considered to have occurred between 2-4 Ma, with the Cape Egmont Fault Zone the main conduit. Timing is constrained by lack of hydrothermal alteration of the hydrocarbons (indicating migration took place after volcanic activity ceased) and age of the seal (i.e., Giant Foresets Formation; Reed, 1992). In a more recent study focused to the south and west of the graben area, McAlpine (2000) suggests that because of overpressures developed as a result of the Giant Foresets Formation, migration might be more complex than previously thought (see following section). This aspect needs to be looked at in considerably more detail in future studies.
- The Giant Foresets Formation is a mud-dominated depositional system, with only thin sandstone units, which often appear disconnected, possibly seriously limiting the migration of hydrocarbons. While topsets are characterised by coarser lithologies, these are probably too young (Nukumaruan to Castlecliffian in age) to form significant traps.
- A primary factor in the migration of hydrocarbons up through the stratigraphic column is the location and timing of movement on faults. Faults can act as conduits or seals to migration, dependant on whether movement occurs during or after migration. Thus faults, both in their relationship to traps, and the timing of movement relative to maturation and migration of hydrocarbons, are a crucial element in evaluating the hydrocarbon prospectivity of the Giant Foresets Formation.
- The Ariki Formation and equivalent aged marly units provide an effective seal across much of the Western Platform and parts of the Northern Graben. Development of any prospect within the Giant Foresets Formation must account for the distribution of these marly units, particularly in conjunction with the location of major fault networks, and whether faults provide a conduit through these units.
- Traps are more likely to be stratigraphic rather than structural. No major structural elements, other than the graben bounding faults, are present within the foresets. Rather, reservoirs may include such architectural elements as (major) channel-fill sands (see e.g., Fig. 5.20), sands within slope channel or overbank deposits, and slope-disconnected basin floor fans, all of which display *some* inter-connectedness (e.g., Fig. 5.25). Characterisation in terms of key events such as generation and charge versus formation of traps and seals, as well as the physical properties of the reservoir rock itself, is vitally important in assessing these elements as viable reservoirs.

For Miocene and younger (potential) reservoirs, seal is provided by a number of stratigraphic units, including the voluminous and muddy Manganui Formation, the marly and condensed Ariki Formation, and the widespread and mud-dominated Giant Foresets Formation. As with reservoirs, emplacement of seal lithologies relative to maturation and expulsion of hydrocarbons needs to be fully understood, as well as the physical properties of the seal itself (e.g., porosity and permeability). For example, Bennett et al. (1992) suggest that the seal above the Kora structure has permeability problems (i.e., is not sufficiently impermeable), and was not thick enough at the time of charge of the reservoir to contain a hydrocarbon column more than 40-120 m thick. Implications from this are that the Giant Foresets Formation may only provide sufficient and effective seal were it is at its thickest (> 2 km).

7.5.2 Thermal regime and implications for maturation and migration

The most significant contribution that the Giant Foresets Formation has had on petroleum systems in the Taranaki Basin is the cumulative effect that rapid and substantial accumulation has had on the underlying thermal regime. Over the last half a decade, detailed modelling has been undertaken in order to understand more fully the thermal characteristics of Taranaki Basin (e.g., Allis and Funnell, 1993; Armstrong et al., 1996, 1998; Funnell et al., 1996).

Thermal gradients across the Taranaki Basin are attributed to basal heat flow and to the thermal effects of erosion and sedimentation, which respectively elevate and depress surface heat flow. Thermal gradients range from 22-33^{o} C/km, with an average gradient of 29^{o} C/km. In the Western Stable Platform region, temperatures are considered to be at a maximum, due mainly to continuous subsidence and lack of uplift and erosion (Armstrong et al., 1994). Hydrocarbon generation and expulsion in Taranaki Basin may have been as early as the early Paleocene in areas where source rocks were buried to depths greater than 2.5 km prior to 60 Ma. However, for wells in the Western Stable Platform region, most potential source rocks are considered to be immature or have just reached expulsion maturity in the southwestern parts of the basin within the last 1 million years (Thrasher, 1992; Armstrong et al., 1994, 1996). Thrasher (1992) indicates that burial depths in the Taranaki Basin are greater than generally required for hydrocarbon generation (up to 5.5-6 km), and only the lower Late Cretaceous coal measures are routinely buried deeper than 5 km. Upper Late Cretaceous transgressive marine and organic-rich coastal plain sediments are not considered by Thrasher (1992) to be significant in the generation of petroleum, because significant volumes do not appear to have been buried to significant depths.

More recent studies (e.g., McAlpine, 2000; Stagpoole, 2000) have modelled the effect that rapid sedimentation of the prograding continental wedge has had on maturation and migration of hydrocarbons. McAlpine (2000) identified compaction-disequilibrium overpressures that have developed within a thick Eocene shale interval on the Western Stable Platform (northwest of the Maui field) as a result of rapid sedimentation during the Miocene to Recent. This accelerated thermal maturity of source rocks, and produced a more complex migration history than previously thought. As well as indicating the existence of mature, oil-prone kitchens in this region, the model also illustrated that secondary migration was affected by hydrodynamic effects set up beneath the prograding foresets, with overpressures greatest at the front of the prograding wedge (Stagpoole, 2000). Fluid flow (groundwater and hydrocarbons) have been modelled as initially migrating in the direction of progradation, but reversing in direction in the Late Pliocene to flow towards the Taranaki Peninsula (McAlpine, 2000; Stagpoole, 2000). McAlpine (2000) suggests that, as a result of this modelling, maturity levels and oil proneness on the Western Stable Platform may have been underestimated, but stresses the complexity associated with oil expulsion, migration and trapping. In particular, the overpressures formed beneath Oligocene limestone and marls have largely prevented vertical migration of hydrocarbons, except via the Cape Egmont Fault Zone, and the hydrodynamic conditions beneath the Giant Foresets Formation would inhibit migration to the west of down-faulted kitchens.

The maturation history for potential source rocks in the Northern Graben is also intimately linked to the overburden provided by Miocene to Recent sediments, including the Mangaa Formation and Giant Foresets Formation. Much of the overburden required for maturation of hydrocarbons (e.g., 5-5.4 km estimated to reach the vitrinite reflectance approximated for Kora-1 oils at the time of generation; Bennett et al., 1992) has only accumulated over the last 5 Ma following deposition of the Pliocene to Pleistocene sedimentary column. Given the *cumulative* thickness of post Miocene sediment (see Figs. 5.9a-d), inferred source rocks and their distribution (refer to Thrasher and Cahill, 1990; Thrasher, 1992), and relationship of known accumulations to major structural trends (i.e. Cape Egmont Fault Zone; Reed, 1992), oil-prone kitchens within the study area are most likely to exist along the central axis of the Northern Graben.

7.5.3 Summary of hydrocarbon prospectivity

Numerous hydrocarbon shows and a single sub-commercial accumulation of hydrocarbons indicate one significant fact – that there is or has been migration of hydrocarbons through the sedimentary column within the study area. As Bennett et al. (1992, pg. 23) state, 'The essential ingredients necessary for the generation, migration and entrapment of hydrocarbons are present in the northern Taranaki Basin'. Recent exploration focusing on the petroleum systems of northern Taranaki Basin, with emphasis on the Northern Graben (e.g., Thrasher et al., 2002), suggest that the potential exists for commercial accumulations of hydrocarbons, predominantly in Neogene structural traps or associated with volcanic centres. However, while bright spots on seismic reflection profiles indicate the presence of hydrocarbons within the Giant Foresets Formation, the probability of large accumulations existing in this formation is low. Nevertheless, the Giant Foresets Formation has contributed substantially to the thermal and hydrodynamic conditions at deeper levels, and as such needs to be investigated further to understand more accurately the nature of underlying petroleum systems.

Chapter 8: Summary and Conclusions

Chapter 8: Summary and Conclusions

8.1 Introduction

This chapter is intended to provide a brief summary of the objectives and conclusions reached in each chapter of the thesis.

8.2 Chapter 1

Chapter 1 introduces the geological occurrence of the Giant Foresets Formation and outlines the reasons for undertaking this study. It defines the thesis objectives and provides an outline of each of the chapters.

8.3 Chapter 2

The late Neogene succession within Taranaki Basin is put into a regional context in this chapter. The evolution of Taranaki Basin, with particular emphasis on northern Taranaki Basin, is reviewed, and the stratigraphy, provenance and relationship of late Neogene units is examined. A brief discussion on the stratigraphic links between Wanganui, King Country and Taranaki Basins are also given.

8.4 Chapter 3

Detailed sampling and foraminiferal analysis of four well sections, in combination with previously published and unpublished data for other well sections, has enabled revision of the biostratigraphy of northern Taranaki Basin. In particular, the positions of New Zealand Stage boundaries have been more tightly constrained, and a chronostratigraphic framework for northern Taranaki Basin, making use of seismic correlation between well sections, has been achieved. The identification and characterisation of condensed sections (e.g., the Ariki Formation) in northern Taranaki Basin has provided new insights into the timing of major phases of volcaniclastic and siliciclastic sedimentation and how these relate to the progradation of continental margin successions versus volcanism in Taranaki Basin and related basins (Wanganui and King Country). The conclusions of Chapter 3 are as follows:

Conclusions

- Revision of the biostratigraphy of Arawa-1, Ariki-1, Kora-1, and Wainui-1, and a review of the latest version of the biostratigraphy of Mangaa-1, has revealed mixed results in terms of the effectiveness of more detailed biostratigraphic evaluation of well sections. Resolution of biostratigraphic stages have been increased dramatically at sites that have remained at deeper water depths, and which contain abundant planktic foraminifera, but more detailed sampling does little to refine biostratigraphic boundaries at sites that shallowed earlier due to progradation of the continental margin and where planktic foraminifera are not common.
- The lower boundary of the Giant Foresets Formation is often associated with an unconformity that encompasses the Miocene-Pliocene boundary. The unconformity embraces an increasingly older and longer-lived time span to the north, while the base of the Giant Foresets Formation also becomes increasing younger to the north.
- While the age of the base of the Giant Foresets Formations is diachronous, it generally falls within the mid to Late Opoitian interval. At Awatea-1 and Mangaa-1, the base of the formation is Waipipian in age. At these two sites, the Giant Foresets Formation is underlain by the Opoitian-Waipipian aged Mangaa Formation, which grades conformably into the Giant Foresets Formation. At Arawa-1, and Taimana-1, Manganui Formation sediments grade conformably into Giant Foresets Formation sediments. At Taimana-1, a mid Opoitian age for the base of the Giant Foresets Formation is inferred, while at Arawa-1, a Waipipian age is interpreted (seismic evidence suggests that the boundary could be placed lower, within the Opoitian Stage).
- The Miocene-Pliocene unconformity is expressed differently across northern Taranaki Basin. In the northwestern regions of the Western Stable Platform and within the Northern Graben, the unconformity is associated with a highly condensed unit (Ariki Formation) or equivalent aged marl and marly siltstone, and is associated with terrigenous sediment starvation, although may be modified in places by local topography. Inability to recognise certain intervals of time above or below the Ariki Formation is attributed to sampling resolution of well cuttings, and thus the unconformity associated with these condensed intervals is better termed a paraconformity. South of the study area (southern Taranakı Basin) and east across the Turi Fault Zone, unconformity development is associated with uplift and erosion of

the footwall of the fault zone. Along the central axis of the Northern Graben, Pliocene sediment rests unconformably over Miocene volcanic massifs.

8.5 Chapter 4

The application of wireline logging interpretative methods to the Late Miocene to Pleistocene succession of eleven well sections drilled in northern Taranaki Basin has enabled the description and characterisation of the Manganui, Mohakatino, Mount Messenger, Urenui, Ariki, Mangaa and Giant Foresets Formations. Emphasis is placed on the identification of wireline facies, based on facies motifs, that can be related to particular depositional systems and depositional mechanisms. A fence diagram linking these eleven wells shows the distribution in time and space of wireline facies. This chapter also introduces the concept of cyclicity through the Giant Foresets Formation, and suggests that the recognition of sequences on the basis of wireline motif alone is not realistic. A number of conclusions have been made:

Conclusions

- Sedimentation patterns in northern (offshore) Taranaki Basin are intimately linked to contemporaneous deposition in Wanganui and King Country Basins. The continental margin associated with the Whangamomona Group only had limited progradation into Taranaki Basin during the Middle Miocene to Early Pliocene, evidenced by thick accumulations of basin floor and slope deposits along the eastern margin of the basin (northern Taranaki coastal outcrop and nearshore wells), and the predominance of hemipelagic sedimentation elsewhere in northern Taranaki Basin.
- The Ariki Formation is a significant marker horizon in offshore northern Taranaki Basin. Based on its wireline motif, time-equivalents of this formation may be more widespread than previously recognised. This formation is particularly relevant as it is related to at least three important events: (i) the limited extent of progradation of the Whangamomona Group foreset front, (ii) initiation of extension of the Northern Graben, and (iii) the timing of the Tangahoe pulldown in Wanganui Basin.
- The wireline motif of the Mangaa Formation is, like the lower part of the Mount Messenger Formation, characteristic of basin floor fan sedimentation. However, the Mangaa Formation appears to be extremely restricted in its extent, indicating that the Northern Graben acted as a sink for sediment, probably initially sourced from contemporaneous uplift to the east.

- It is difficult to characterise the Giant Foresets Formation on the basis of geophysical wireline motif. Often only very subtle changes (in wireline motif) are noted uphole, although GR (and occasionally sonic and density logs) tend to display an overall coarsening-upward trend. At most sites, the upper few hundred metres of each logged well section are distinctly different, reflecting the coarser deposits of the shelf facies (topsets).
- Six facies based on wireline log motifs have been mapped in the northern Taranaki Basin: (i) hemipelagic facies, (ii) basin floor fan facies, (iii) slope fan facies, (iv) shelf facies, (v) slumped facies, and (vi) condensed facies.
- The Miocene is dominated by hemipelagic (basinal) facies, with interfingering of coarser lithologies of the volcaniclastic basin floor fan facies (Bff2), and volcaniclastic Sf3 sediments in the north and east of the study area.
- Pliocene basin facies (bottomsets) are also dominated by hemipelagic sedimentation (other than the Mangaa Formation), although both the sporadic occurrence of thin basin floor fan facies and the increasing prevalence of Sf2 (distal fan sedimentation?) are suggestive of the increasing proximity of the advancing foreset front.
- Arrival of the foreset front at any one site is documented by the frequent presence of Sf1 and slumped facies. The latter facies are more often than not associated with degradational foresets, and a change in the physical properties of the clay matrix.
- While wireline logs are an effective tool for identifying depositional environments, and in shelfal settings can be used effectively for identifying sequences and systems tracts, serious limitations arise in deeper water situations. Within offshore northern Taranaki Basin sediments, bathyal water depths, which dominated until the Pleistocene, and distance from the depositional shoreline, resulted in a predominance of muddy hemipelagic facies. Though the Giant Foresets Formation does display fine scale coarsening- or fining-upward cyclical packages (Sf1, Sf2, and Shf1), resolution of the logging tools, and lack of well-defined biostratigraphic dating, means that ambiguity is inherent in any interpretation.
- Three levels of cyclicity can be recognised: 2nd-order (several millions of years) related to progradation of the Whangamomona Group and Rangitikei Supergroup, 4th-order (400 ka) cycles, and ?5th-order (100 ka) cycles (possibly related to glacio-eustatic controls). Neither the 4th or 5th-order cycles can be consistently correlated between well sites.

8.6 Chapter 5

The seismic characteristics of the Giant Foresets Formation and underlying formations are reviewed in this chapter. The Giant Foresets Formation in particular is discussed in detail, and over 70 discrete seismic units have been mapped through this succession in northern Taranaki Basin. Several of these seismic units have been mapped across a seismic grid, and are used to help interpret the progradation of the continental wedge across the study region. The significance of the Mangaa and Ariki Formations, precursors to Giant Foresets Formation deposition, are also discussed in more detail in this chapter. Identification of a number of seismically-defined architectural elements has enabled the identification of several seismic sequences on seismic reflection profile P95-103. Cyclicity is determined to be in the order of c.400 ka (4^{th}-order). Higher resolution seismic sequences are not able to be delineated on the seismic reflection profiles interpreted in this study. The conclusions reached in this study are as follows:

Conclusions

- The Giant Foresets Formation is well imaged on seismic reflection profiles. Seismic units can be divided into topset, foreset (progradational and degradational) and bottomset beds (after Beggs, 1990). While underlying formations are generally characterised by parallel to subparallel reflectors, the Giant Foresets Formation is characterised by a diversity of reflector characteristics, suggesting several depositional mechanisms.
- Progradation into the study area initially occurred as a series of individual fan lobes, probably fed from a single point source. The distribution pattern of sediment associated with the advancing continental margin gradually changed to a more linear relationship, reflecting increasing shelf width, increasing sediment flux, and decreasing influence of local tectonics. Structure-contour, isopach, and shelf break maps demonstrate the overall north-northwest translation of depocentres, and illustrate how the depositional loci changed throughout the Plio-Pleistocene.
- Classic sigmoidal-shaped progradational seismic units of the Giant Foresets Formation typically comprise a number of architectural elements that reflect sediment flux and the dominant lithology. Older (Waipipian-Mangapanian) seismic units are characterised by chaotic, internally disrupted to reflection free configurations consistent with slumping. Progressive oversteepening led to slope failure, while

absence of channels and the presence of failure chutes characterise a mud-rich line-source slope apron system with an increasing sediment supply. Younger progradational units are distinguished by numerous small channel-levee systems on the lower and middle slope, and larger meandering channel systems on the shelf and upper slope. Basin floor fans are characterised by low-relief mounds and/or gull-wing geometries of channel-levee systems, or sub-parallel reflectors. These characteristics are more typical of a mixed mud/sand-rich depositional system.

- Mapping of the time-rich Ariki Formation over the northern Western Stable Platform, and identification of two seismically unrelated but age-equivalent condensed intervals within the Northern Graben indicate that sediment starvation, possibly accentuated by several warming trends/flooding events in the Late Miocene to Early Pliocene, impacted substantially on the evolution of the sedimentary succession in the study area. Paleobathymetric relief may have been locally important in extending the duration of sediment starvation.
- The presence of voluminous siliciclastic deposits in the Northern Graben during the mid Opoitian to Waipipian (Mangaa Formation), and the paucity of equivalent aged sediment elsewhere in the study area, indicates that the Northern Graben was an efficient and important depositional sink during this period. Channel orientation and interpreted sediment pathways indicate that the Northern Graben continued to be a focus for sediment deposition during the Late Pliocene and Pleistocene.
- Identification of architectural elements within the Giant Foresets Formation has allowed the preliminary identification of several 4^{th}-order (400 k.y.) sequences. These sequences contain architectural elements that are linked to mechanisms and depositional styles that are more frequently observed with lowering and low relative sea level (RST and LST), such as basin floor (lower slope) fans, and channel-levee complexes. Basinward of the shelf margin, the HST of each sequence is represented by a bright, high amplitude and laterally continuous reflector that rolls over into shelf deposits. Transgressive systems tract components are not recognised basinward of the shelf margin, and are incorporated within the HST landward of the shelf margin. The HST (and TST?) is time-rich relative to the RST/LST, resulting in a highly asymmetrical sequence architecture.
- The predominant factor controlling the seismic character of the Giant Foresets Formation during the Late Opoitian to Mangapanian appears to be sediment flux and lithology over relative sea level change. However, during the Nukumaruan to

Castlecliffian, decreasing accommodation and rapidly shallowing paleobathymetry as a result of accelerated progradation meant that changes in relative sea level increasingly influenced stratal architecture.

8.7 Chapter 6

This chapter investigates the (foraminiferal) paleoenvironmental signatures for four well sections (Arawa-1, Ariki-1, Kora-1 and Wainui-1) interpreted from detailed analyses of well cuttings. Statistical analyses of both planktic and benthic foraminiferal species were used to compare well sections and determine their paleobathymetry and watermass conditions through the Plio-Pleistocene section. In addition, the influence of, and relationship between, sea level change and benthic assemblages has been investigated. The cyclicity evident in the Giant Foresets Formation is investigated further in this chapter. Integration of wireline, seismic, and paleoenvironmental information has revealed up to thirteen (4^{th}-order) sequences, although many of these cannot be directly correlated between well sections.

Conclusions

- Detailed studies of benthic and planktic foraminifera are essential in reconstructing paleoenvironmental trends through time, and establishing paleobathymetry and watermass conditions.
- Paleobathymetric curves for Arawa-1, Ariki-1, Kora-1, Mangaa-1, and Wainui-1 show a progressive shallowing from the Late Miocene to Pleistocene, with all sites reaching shelfal depths by the Late Nukumaruan to Castlecliffian.
- A sharp decrease in planktic percentage and an apparently dramatic change in watermass conditions is observed between latest Miocene/Opoitian strata and Waipipian strata. Paleobathymetric curves indicate that water depths did not change as significantly as changing watermass conditions would imply. Thus the initially dramatic change may be related to deflection of local current systems away from northern Taranaki Basin by the advancing progradational front, affecting surface water nutrient supply and therefore planktic percentage. Increased suspended sediment load associated with increasing sediment flux through the Late Pliocene and Pleistocene could account for an observed planktic increase during this interval at Ariki-1.

- Benthic foraminiferal data are particularly useful in demonstrating sediment pathways across the study area; benthic assemblages, even at the deepest sites, are often characterised by the postmortem mixing of shallow water faunas, indicating that at times there was effective sediment transport across the shelf to the slope and basin floor.
- The integration of benthic foraminiferal data, including faunal composition, diversity indices, and species dominance, with watermass conditions and textural curves, can provide a proxy for indicating periods of lowered relative sea level in the absence of other criteria. However, inherent difficulties with cuttings samples and sample resolution problems mean that not all paleoenvironmental signals will be identified.
- The integration of seismic, wireline and paleoenvironmental data has enabled the recognition of ten or more 4^{th}-order sequences in northern Taranaki Basin. However, correlation of sequences, predominantly because of the wide spacing of wells, is in reality only possible using seismic reflection profiles. More detailed research on this aspect of the Giant Foresets Formation needs to be undertaken to construct a more robust sequence stratigraphic model.

8.8 Chapter 7

This chapter reports the results of backstripping and decompaction of eleven well sections in northern Taranaki Basin, as well as backstripping, decompaction, and palinspastic restoration of a complete seismic reflection profile (P95-158). The results from these backstripped wells and seismic line were integrated with information interpreted from previous chapters to summarise the geological evolution of the Giant Foresets Formation through the Pliocene and Pleistocene in a series of paleogeographic maps. This chapter also summarised the hydrocarbon systems of northern Taranaki Basin, and discussed the role that the Giant Foresets Formation has played in maturation and migration of hydrocarbons.

Summary of the evolution of the Giant Foresets Formation, northern Taranaki Basin

The Giant Foresets Formation is the latest phase of a major 1^{st}-order transgressive-regressive megacycle that began in the Late Cretaceous and continues to the present day. Kamp et al. (2002) identified four 2^{nd}-order megacycles in Wanganui and King Country Basins superimposed on the 1^{st}-order transgressive-regressive cycle, the youngest two of which are the

Whangamomona Group/megacycle and the Rangitikei Supergroup/megacycle. The Giant Foresets Formation represents the Rangitikei Supergroup in northern Taranaki Basin.

The Whangamomona Group incorporates the Otunui, Mount Messenger, Urenui, Kiore, and Matemateaonga Formations. The progradational wedge associated with this Group had limited extent in northern Taranaki Basin, reaching only the eastern margin of the basin. This contributed in part to the development of a significant paraconformity across much of the Western Stable Platform (Ariki Formation), and other thinner condensed horizons in the subsiding Northern Graben. However, the rapid subsidence during the late Early Pliocene of the Wanganui Basin depocentre (Tangahoe pulldown), which acted to divert and trap sediment derived from erosion of the Southern Alps, is considered a significant factor in the development of the Ariki Formation, and the general sediment starvation observed during this time in northern Taranaki Basin.

Progradation of the Rangitikei Supergroup began with deposition of the Tangahoe Mudstone, and the megasequence migrated northward as two fronts, one through Wanganui Basin and into King Country Basin, and a second west of the Patea-Tongaporutu High and into northern Taranaki Basin (Kamp et al., 2002). The Giant Foresets Formation is the stratigraphic equivalent of the Tangahoe Mudstone and younger progradational shelf deposits in Wanganui Basin.

During the mid to Late Opoitian, northern Taranaki Basin was characterised by sediment starvation in the northwestern region of the Western Stable Platform, and voluminous siliciclastic sedimentation in the newly created accommodation in the Northern Graben (Mangaa Formation). The source for Mangaa Formation sediments was probably erosion of Whangamomona Group strata in King Country Basin. The leading edge of the progradational wedge (Giant Foresets Formation) began to extend into the southern regions of the study area during the Opoitian. It is possible that the advancing foreset front acted to deflect local current systems away from some sites, leading to a marked decrease in oceanicity and planktic percentages ahead of the decrease in water depth.

By the Waipipian, the continental margin had prograded further into the study area, and water depths began to shallow, although they still remained predominantly bathyal. Sedimentation rates increased through the Waipipian and Mangapanian, consistent with an acceleration in the rate of erosion of the Southern Alps. Several channel systems fed sediment to distal sites, and mixing of shallow and deep water benthic foraminifera suggests there was efficient communication between the shelf and slope. The southernmost region attained outermost bathyal to upper shelf conditions by the Late Waipipian, although the northern region had rapidly shallowed from mid/lower bathyal to upper bathyal by the Early Nukumaruan.

The continental margin advanced rapidly during the Early Nukumaruan, and the leading edge of the foreset front reached the northernmost region of the study area by the Late Nukumaruan to Castlecliffian. Several channel networks provided a route by which sediment was transported from the shelf to the slope. The most voluminous Nukumaruan sedimentation occurred in the northern regions of the study area, associated with the progradational wedge; to the south, only thin shelfal deposits were accumulating. Sedimentation rates dropped off markedly after passage of the progradational front.

The youngest foreset units to be deposited (degradational foresets) have a different sediment texture to older foreset units, which may reflect an increasing contribution from sediment derived from erosion of North Island's Herangi Range, and possibly also a volcaniclastic Taupo Volcanic Zone source. The focus of sedimentation has now moved north and west of the study area.

References

References

Abbott, S.T., Carter, R.M., 1997: Macrofossil associations from Mid-Pleistocene cyclothems, Castlecliff section, New Zealand: Implications for sequence stratigraphy. *Palios, 12:* 188-210.

Adams, E.W., Schlager, W., Wattle, E., 1998: Submarine slopes with an exponential curvature. *Sedimentary Geology, 117:* 135-141.

Adams, E.W., Schlager, W., 2000: Basic types of submarine slope curvature. *Journal of Sedimentary Research, 70, no. 4:* 814-828.

Ahmadi, A.M., Coe, A.L., 1998: Methods for simulating natural gamma ray and density wireline logs from measurements on outcrop exposures and samples: examples from the Upper Jurassic, England. *In* Harvey, P.K., & Lovell, M.A. (eds) Core-Log Integration. *Geological Society, London, Special Publications 136:* 65-80.

Allis, R., Funnell, R., 1993: Oil and gas generation beneath onshore Taranaki Basin: implications of varying thermal regime. *Petroleum exploration in New Zealand news, 38:* 10-17.

Altenbach, 1988: Deep-sea benthic foraminifera and flux rate of organic carbon (Résumé - Abstract). *In* “Benthos’ 1986”. Rev. Paléobiol. Vol. Spécial, 2, Partié 2: 719-720.

Alve, E., 1999: Colonization of new habitats by benthic foraminifera: a review. *Earth-Science Reviews, 46*: 167-185.

Alve, E., Murray, J.W., 1997: High benthic fertility and taphonomy of foraminifera: a case study of the Skagerrak, North Sea. *Marine Micropaleontology, 31:* 157-175.

ARCO Petroleum NZ Inc., 1985: Final Well Report, Kora-4, PPL38447. Ministry of Commerce New Zealand, Unpublished Petroleum Report 1443.

ARCO Petroleum NZ Inc., 1988: Final Well Report, Kora-1, Kora-1A, PPL38447. Ministry of Commerce New Zealand, Unpublished Petroleum Report 1374.

ARCO Petroleum NZ Inc., 1992: Arawa-1 Final Well Report PPL38436. Ministry of Commerce New Zealand, Unpublished Petroleum Report 1824.

Armstrong, P.A., Chapman, D.S., Funnell, R.H., Allis, R.G., Kamp, P.J.J., 1994: Thermal state, thermal modelling, and hydrocarbon generation in the Taranaki Basin, New Zealand. *In* 1994 New Zealand Petroleum Conference proceedings: 289-307. Ministry of Commerce, Wellington.

Armstrong, P.A., Chapman, D.S., Funnell, R.H., Allis, R.G., Kamp, P.J.J., 1996: Thermal modelling and hydrocarbon generation in an active-margin basin: Taranaki Basin, New Zealand. *American Association of Petroleum Geologists Bulletin, vol. 80, no. 8:* 1216-1241.

Armstrong, P.A., Allis, R.G., Funnell, R.H., Chapman, D.S., 1998: Late Neogene exhumation patterns in Taranaki Basin (New Zealand); Evidence from offset porosity-depth trends. *Journal of Geophysical Research, vol. 103, no. B12:* 30269-30282.

Asquith, G.B., 1982: Basic well log analysis for geologists. *American Association of Petroleum Geologist, Methods in Exploration Series*. 216 p.

Bally, A.W., (ed) 1987: AAPG Atlas of Seismic Stratigraphy, 3 volumes. *American Association of Petroleum Geologists studies in geology no. 27.* American Association of Petroleum Geologists, Tulsa.

Barker, R. Wright, 1960: Taxonomic Notes on the species figured by H.B. Brady in his report on the foraminifera dredged by H.M.S *Challenger* during the years 1873-1876. *Society of Economic Paleontologists and Mineralogists Special Publication No. 9.*

Beals, E.W., 1984: Bray-curtis ordination: an effective strategy for analysis of multivariate ecological data. *In* Macfadyen, A., Ford, E.D. (eds) *Advances in Ecological Research*. Academic press, London. 1-55.

Beggs, J.M., 1990: Seismic stratigraphy of the Plio-Pleistocene Giant Foresets, Western Platform, Taranaki Basin. *In* 1989 NZ Oil Exploration Conference proceedings: 201-207. Ministry of Commerce, Wellington.

Bennett, C., Gregg, R., King, P.R., 1992: Petroleum geology of the Taranaki Basin, with emphasis on the north-eastern quadrant. *Petroleum exploration in New Zealand news, 32:* 15-25.

Berg, O.R., and Woolverton, D.G. (eds) 1985: Seismic Stratigraphy II: An integrated approach to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 39.* 256p.

Berger, W.H., Diester-Haass, L., 1988: Paleoproductivity: the benthic/planktonic ratio in foraminifera as a productivity index. *Marine Geology, 81*: 15-25.

Berggren, W.A., 1978: Marine micropaleontology: an introduction. *In* Haq B.U., Boersma, A. (eds) *Introduction to Marine Micropaleontology.* Elsevier, New York. 1-17.

Bergman, S.C., Atkinson, C.D., Talbot, J., Gordon, T.L., 1990: Nature and reservoir potential of Miocene sedimentary and volcanic rocks, Western North Island, NZ, PPL 38449. New Zealand Unpublished Petroleum Report 1581. Ministry of Commerce, Wellington.

Bergman, S.C., Talbot, J., Thompson, P.R., 1992: The Kora Miocene submarine andesite stratovolcano hydrocarbon reservoir, Northern Taranaki Basin, New Zealand. *In* 1991 New Zealand Oil Exploration Conference proceedings: 178-206. Ministry of Commerce, Wellington.

Betzler, C., Reijmer, J.J.G., Bernet, K., Eberli, G.P., Anselmettis, F.S., 1999: Sedimentary patterns and geometries of the Bahamian outer carbonate ramp (Miocene-Lower Pliocene, Great Bahama Bank). *Sedimentology, 46:* 1127-1143.

Boltovskoy and Wright, 1976: Recent foraminifera. Junk, The Hague. 515 p.

Bosario, I., 1981: Seismic stratigraphy of Egmont Terrace, offshore Taranaki. Unpublished M.Sc. thesis, Victoria University of Wellington, New Zealand.

Bouma, A.H., 2000a: Coarse-grained and fine-grained turbidite systems as end member models: applicability and dangers. *Marine and Petroleum Geology, 17*: 137-143.

Bouma, A.H., 2000b: Fine-grained, mud-rich turbidite systems: model and comparison with coarse-grained, sand-rich systems. *In* Bouma, A.H., Stone, C.G. (eds) Fine-grained turbidite systems. *AAPG Memoir 72/SEPM Special Publication no. 68:* 9-20.

Bradford, J.M., Roberts, P.E., 1978: Distribution of reactive phosphorus and plankton in relation to upwelling and surface circulation around New Zealand. *New Zealand Journal of Marine and Freshwater Research, 12:* 1-15.

Brown, L.F., Jr., Fisher, W.L., 1980: Seismic stratigraphic interpretation and petroleum exploration. *American Association of Petroleum Geologists Course Notes 16.* 181 p.

Browne, G.H., McAlpine, A., King, P.R., 1996: An outcrop study of bed thickness, continuity and permeability in reservoir facies of the Mt Messenger Formation, North Taranaki. *In* 1996 New Zealand Petroleum Conference proceedings: 154-163. Ministry of Commerce, Wellington.

Browne, G.H., Slatt, R.M., 1997: Thin-bedded slope fan (channel-levee) deposits from New Zealand: An outcrop analog for reservoirs in the Gulf of Mexico. *Gulf Coast Association of Geological Societies Transactions, XLVII:* 75-86.

Browne, G.H., Slatt, R.M., 2002: Outcrop and behind-outcrop characterization of a late Miocene slope fan system, Mt. Messenger Formation, New Zealand. *American Association of Petroleum Geologists Bulletin, vol. 86, no. 5:* 841-862.

Browne, G.H., Slatt, R.M., King, P.R., 2000: Contrasting styles of basin-floor fan and slope fan deposition: Mount Messenger Formation, New Zealand. *In* Bouma, A.H., Stone, C.G. (eds) Fine-grained turbidite systems. *AAPG Memoir 72/SEPM Special Publication no. 68:* 143-152.

Bussell, M.R., 1994: Seismic interpretation of the Moki Formation on the Maui 3D survey, Taranaki Basin. *In* 1994 New Zealand Petroleum Conference proceedings: 240-255. Ministry of Commerce, Wellington.

Buzas, M.A., Gibson, T.G., 1969: Species Diversity: Benthic foraminifera in the western North Atlantic. *Science, 163:* 72-75.

Cant, D.J., 1992: Subsurface facies analysis. *In* Walker, R.G., and James, N.P. (eds) Facies models: response to sea level change. *Geological Association of Canada*: 297-310.

CANZ, 1997: New Zealand Region Bathymetry, 1:4000 000. NIWA Chart Miscellaneous Series No. 73.

Chappell, J., Shackleton, N.J., 1986: Oxygen isotopes and sea level. *Nature, 324*: 137-140.

Chappell, J., Omura, A., Esat, T., McCulloch, M., Pandolfi, J., Ota, Y., Pillans, B., 1996: Reconciliation of late Quaternary sea levels derived from coral terraces at Juon Peninsula with deep sea oxygen isotope records. *Earth and Planetary Science Letters, 141*: 227-236.

Coffeen, J.A., 1986: *Seismic Exploration Fundamentals* (2nd Edition). PennWell Publishing Company, Tulsa. 347 p.

Coleman, J.L., Browne, G.H., King, P.R., Slatt, R.M., Spang, R.J., Williams, E.T., Clemenceau, G.R., 2000a: The inter-relationships of scales of heterogeneity in subsurface, deep water E&P projects – lessons learned from the Mount Messenger Formation (Miocene), Taranaki Basin, New Zealand. GCSSEPM Foundation 20th Annual Research Conference. Deep-water Reservoirs of the World, December 306, 2000. 263-281.

Coleman, J.M., Prior, D.B., 1988: Mass wasting on continental margins. *In* Wetherill, G.W., Albee, A.L., Stehli, F.G. (eds) *Annual Review of Earth and Planetary Sciences, 16*: 101-119.

Coleman J.L., Sheppard III, F.C., Jones, T.K., 2000: Seismic resolution of submarine channel architecture as indicated by outcrop analogs. *In* Bouma, A.H., Stone, C.G. (eds) Fine-grained turbidite systems. *AAPG Memoir 72/SEPM Special Publication no. 68:* 119-126.

Cooke, P.J., 2002: *Aspects of Neogene palaeoceanography in the southern Tasman Sea (DSDP Site 593).* Unpublished PhD thesis, University of Waikato. 334 p.

Corliss, B.H., 1991: Morphology and microhabitat preferences of benthic foraminifera from the northwest Atlantic Ocean. *Marine Micropaleontology, 17*: 195-236.

Crews, J.R., Weimer, P., Pulham, A.J., Waterman, A.S., 2000: Integrated approach to condensed section identification in intraslope basins, Pliocene-Pleistocene, northern Gulf of Mexico. *American Association of Petroleum Geologists Bulletin, vol. 84, no. 10:* 1519-1536.

Cronin, B.T., Hurst, A., Celik, H., Türkmen, I., 2000: Superb exposure of a channel, levee and overbank complex in an ancient deep-water slope environment. *Sedimentary Geology, 132*: 205-216.

Cross, T.A., Lessenger, M.A., 1988: Seismic stratigraphy. *In* Wetherill, G.W., Albee, A.L., Stehli, F.G. (eds) *Annual Review of Earth and Planetary Sciences, 16*: 319-354.

Crowley, J., Crocker, S.J., 1989: Well completion report, Okoki-1, PPL 38438. Ministry of Commerce New Zealand Unpublished Petroleum Report 1495.

Crundwell, M.P., Scott, G.H., Strong, C.P., 1992: Biostratigraphy of Arawa-1 offshore petroleum exploration well North Taranaki Basin. Contract Report No. 1992/18. New Zealand Department of Scientific and Industrial Research (Geology and Geophysics).

Crundwell, M.P., Scott, G.H., Thrasher, G.P., 1994: Calibration of paleobathymetry indicators by integrated seismic and paleontological analysis of foreset sequences, Taranaki Basin, New Zealand. *In* 1994 New Zealand Petroleum Conference proceedings: 169-178. Ministry of Commerce, Wellington.

Dam, G., Sϕnderholm, M., 1994: Lowstand slope channels of the Itilli succession (Maastrichtian-Lower Paleocene), Nuussuaq, West Greenland. *Sedimentary Geology, 94:* 49-71.

Danielsen, M., Michelsen, O., Clausen, O.R., 1997: Oligocene sequence stratigraphy and basin development in the Danish North Sea sector based on log interpretations. *Marine and Petroleum Geology, 14, No. 7/8:* 931-950.

Davis, T.L., 1992: Seismic-stratigraphic models. *In* Walker, R.G. and James, N.P. (eds) Facies models: response to sea level change. *Geological Association of Canada.* 311-317.

Davis, G.H., Reynolds, S.J., 1996: *Structural geology of rocks and regions.* John Wiley, New York. 776 p.

Dawson, E.W., 1992: The marine fauna of New Zealand: Index to the Fauna 1. Protozoa. New Zealand Oceanographic Institute memoir, 99. 368 p.

de Bock, J.F., 1994: Moki Formation, a Miocene reservoir sequence, its facies distribution and source in offshore southern Taranaki Basin. *In* 1994 New Zealand Petroleum Conference proceedings: 155-167. Ministry of Commerce, Wellington.

Denne, R.A., Sen Gupta, B.K., 1991: Association of bathyal foraminifera with water masses in the northwestern Gulf of Mexico. *Marine Micropaleontology, 17:* 173-193.

Diamond Shamrock Oil Co (NZ) International Department Dallas, 1984: Final well report Taimana-1 PPL 38109. Ministry of Commerce Unpublished Petroleum Report 1026.

Doveton, J.H., 1994: Geologic log interpretation. *Society of Economic Paleontologists and Mineralogists Short Course No. 29.* Tulsa, U.S.A.

Eade, J.V., 1967: A checklist of New Zealand foraminifera. New Zealand Department of Scientific and Industrial Research Bulletin 182. 72p.

Elverhøi, a., Norem, H., Andersen, E.S., Dowdeswell, J.A., Fossen, I., Haflidason, H., Kenyon, N.H., Laberg, J.S., King, E.L., Sejrup, H.P., Solheim, A., Vorren, T., 1997: On the origin and flow behaviour of submarine slides on deep-sea fans along the Norwegian-Barents Sea continental margin. *Geo-Marine Letters, 17:* 119-125.

Emery, D, Myers, K.J., (eds) 1996: Sequence stratigraphy. Blackwell Science. Oxford and Northampton, Great Britain. 297 p.

Eschard, R., 2001: Geological factors controlling sediment transport from platform to deep basin: a review. *Marine and Petroleum Geology, 18:* 487-490.

Farre, J.A., McGregor, B.A., Ryan, W.B.F., Robb, J.M., 1983: Breaching the shelfbreak: passage from youthful to mature phase in submarine canyon evolution. *In* Stanley, D.J., Moore, G.T. (eds) The shelfbreak: critical interface on continental margins. *Society of Economic Paleontologists and Mineralogists Special Publication no. 33*: 25-39.

Fisher, R.A., Corbet, A.S., Williams, C.B., 1943: The relation between the number of species and the number of individuals in a random sample of an animal population. *Journal of Animal Ecology, 12:* 42-58.

Forder, S.P., Sissons, B.A., 1992: The Moki C Sands: an example of Mio-Pliocene bathyal fans in the North Taranaki Graben. *In* 1991 New Zealand Oil Exploration Conference proceedings: 155-167. Ministry of Commerce, Wellington.

Fulthorpe, C.S., Austin Jr., J.A., 1998: Anatomy of rapid margin progradation: three-dimensional geometries of Miocene clinoforms, New Jersey margin. *American Association of Petroleum Geologists Bulletin, 82:* 251-273.

Funnell, R., Chapman, D., Allis, R., Armstrong, P., 1996: Thermal state of the Taranaki Basin, New Zealand. *Journal of Geophysical Research, vol. 101, no. B11:* 25 197 – 25 215.

Galloway, W.E., 1989: Genetic stratigraphic sequences in basin analysis: architecture and genesis of flooding surface bounded depositional units. *American Association of Petroleum Geologists Bulletin, 73:* 125-142.

Galloway, W.E., 1998: Siliciclastic slope and base-of-slope depositional systems: component facies, stratigraphic architecture, and classification. *American Association of Petroleum Geologists Bulletin, vol. 82, no. 4:* 569-595.

Gibson, T.G., 1989: Planktonic benthonic foraminiferal ratios: Modern patterns and Tertiary applicability. *Marine Micropaleontology, 15*: 29-52.

Gibson, T.G., Buzas, M.A., 1973: Species diversity: patterns in Modern and Miocene foraminifera of the eastern margin of North America. *Geological Society of America Bulletin, 84*: 217-238.

Glennie, K.W., 1958: The upper Miocene formations of west Taranaki. New Zealand unpublished openfile petroleum report 419. Ministry of Commerce, Wellington.

Gooday, A.J., 1988: A response by benthic foraminifera to the deposition of phytodetritus in the deep sea. *Nature, vol. 332, no. 3:* 70-73.

Gooday, A.J., 1994: The biology of deep-sea foraminifera: A review of some advances and their applications in paleoceanography. *PALAIOS, 9*: 14-31.

Gray, G., Rankin, J., Wellings, J., 1988: Final well report Te Kumi-1, PPL 38449. Ministry of Economic Development, New Zealand. Unpublished Petroleum Report 1386.

Griffin, A.G., 2001: *Late Cenozoic subsurface geology of the Taranaki Peninsula region based on analysis of geophysical well logs.* Unpublished M.Sc. thesis, University of Waikato, New Zealand.

Gupta, A.K., Srinivasan, M.S., 1992: Species diversity of Neogene deep sea benthic foraminifera from Northern Indian Ocean DSDP Sites 214 and 216A. *In* Takayanagi, T., and Saito, T. (eds) *Studies in Benthic Foraminifera: Proceedings of the fourth international symposium on benthic foraminifera, Sendai, 1990.* Tokai University Press, Japan. 249-254.

Hansen, R.J., 1996: *Stratigraphy, sedimentology, and paleomagnetism of a late Miocene succession, Eastern Taranaki Basin Margin.* Unpublished M.Sc. thesis, University of Waikato, New Zealand.

Haq B.U., Boersma, A., (eds) 1978: *Introduction to Marine Micropaleontology.* Elsevier, New York. 376 p.

Haq, B.U., Hardenbol, J., Vail, PR., 1987: Chronology of fluctuating sea-levels since the Triassic. *Science, 235:* 1153-1165.

Haq, B.U., Hardenbol, J., Vail, PR., 1988: Mesozoic and Cenozoic chronostratigraphy and cycles of sea-level change. *In* Wilgus, B.S., Hastings, C.G. St Kendall, Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (eds) Sea-level Changes: an Integrated Approach. *Special Publication, Society of Economic Paleontologists and Mineralogists, Tulsa, 42:* 40-45.

Hay, R.F., 1967: Sheet 7 Taranaki. Geological map of New Zealand 1:250 000. New Zealand Geological Survey. Department of Scientific and Industrial Research, Wellington.

Hayward, B.W., 1984: Foraminiferal biostratigraphy of Wainui-1 Offshore well, West Taranaki. New Zealand Geological Survey report PAL 83.

Hayward, B.W., 1985a: Abundant planktic foraminifera in intertidal sediments, Kawerua, Northland. *Tane, 31:* 1-12.

Hayward, B.W., 1985b: Foraminiferal biostratigraphy of Mangaa-1 offshore well, northwest Taranaki. New Zealand Geological Survey report PAL 104.

Hayward, B.W., 1986a: A guide to paleoenvironmental assessment using New Zealand Cenozoic foraminiferal faunas. New Zealand Geological Survey Report PAL 109. New Zealand Geological Survey DSIR, Lower Hutt, New Zealand.

Hayward, B.W., 1986b: Foraminiferal biostratigraphy and paleobathymetry of Ariki-1 offshore well, northwest Taranaki. New Zealand Geological Survey report PAL 110.

Hayward, B.W., 1987: Paleobathymetry and structural and tectonic history of Cenozoic drillhole sequences in Taranaki Basin. New Zealand Geological Survey report PAL 122.

Hayward, B.W., 1990: Use of foraminiferal data in analysis of Taranaki Basin, New Zealand. *Journal of Foraminiferal Research, vol. 20, no. 1:* 71-83.

Hayward, B.W., Buzas, M.A., 1979: Taxonomy and paleoecology of Early Miocene benthic foraminifera of Northern New Zealand and the North Tasman Sea. *Smithsonian Contributions to Paleobiology, no. 36.* 154 p.

Hayward, B.W., Carter, R., Grenfell, H.R., Hayward, J.J., 2001: Depth distribution of deep-sea foraminifera east of New Zealand, and their potential for improving paleobathymetric assessments of Neogene microfaunas. *New Zealand Journal of Geology and Geophysics, 44:* 555-587.

Hayward, B.W., Grenfell, H., Reid, C., 1997a: Forammininferal associations in Wanganui Bight and Queen Charlotte Sound, New Zealand. *New Zealand journal of marine and Freshwater Research, 31*: 337-365.

Hayward, B.W., Grenfell, H., Reid, C.M., Hayward, K.A., 1999: *Recent New Zealand shallow-water benthic foraminifera: taxonomy, ecologic distribution, biogeography, and use in paleoenvironmental assessment.* Institute of Geological & Nuclear Sciences monograph 21. 264 p. Lower Hutt, New Zealand: Institute of Geological & Nuclear Sciences Limited.

Hayward, B.W., Hollis, C.J. Grenfell, H.R., 1997b: Recent Elphidiidae (Foraminiferida) of the South-west Pacific and fossil Elphidiidae of New Zealand. *Institute of Geological & Nuclear Sciences Monograph 16.* 170p. Lower Hutt, New Zealand: Institute of Geological & Nuclear Sciences Limited.

Hayward, B.W., Neil, H., Carter, R., Grenfell, H.R., Hayward, J.J., 2002: Factors influencing the distribution patterns of Recent deep-sea benthic foramininifera, east of New Zealand, Southwest Pacific Ocean. *Marine Micropaleontology, 46:* 149-176.

Hayward, B.W., Strong, C.P., 1988: Biostratigraphy of Kora-1 offshore well, North Taranaki. New Zealand Geological Survey report PAL 130.

Hayward, B.W., Strong, C.P., 1993: Biostratigraphy of Kora-1 offshore well, North Taranaki. Institute of Geological and Nuclear Sciences science report 93/13.

Hayward, B.W., Wood, R.A., 1989: Computer-generated geohistory plots for Taranaki drillhole sequences. New Zealand Geological Survey report PAL 115.

Hedley, R.H., Hurdle, C.M., Burdett, I.D.J., 1965: A Foraminiferal fauna from the Western continental Shelf, North Island, New Zealand. New Zealand DSIR Bulletin 163. New Zealand Oceanographic Institute Memoir no. 25: 48p

Helland-Hansen, W., Gjelberg, J.G., 1994: Conceptual basis and variability in sequence stratigraphy: a different perspective. *Sedimentary Geology, 92:* 31-52.

Hematite Petroleum (NZ) Ltd., 1970: Mangaa-1 (Offshore). Ministry of Commerce New Zealand Unpublished Petroleum Report 554.

Herzer, R.H., 1995: Seismic stratigraphy of a buried volcanic arc, Northland, New Zealand and implications for Neogene subduction. *Marine and petroleum geology, vol. 12, no. 5:* 511-531.

Holt, W.E., Stern, T.A., 1991: Sediment loading on the Western Platform of the New Zealand continent: Implications for the strength of a continental margin. *Earth and Planetary Science Letters, 107:* 523-538.

Holt, W.E., Stern, T.A., 1994: Subduction, platform subsidence, and foreland thrust loading: The late Tertiary development of Taranaki Basin, New Zealand. *Tectonics, vol. 13, no. 5:* 1068-1092.

Hornibrook, N. de B., 1981: *Globorotalia* (planktic Foraminiferida) in the Late Pliocene and Early Pleistocene of New Zealand. *New Zealand Journal of Geology and Geophysics, 24:* 263-292.

Hornibrook, N. de B., 1982: Late Miocene to Pleistocene *Globorotalia* (Foraminiferida) from DSDP Leg 29, Site 284, Southwest Pacific. *New Zealand Journal of Geology and Geophysics, 25:* 83-99.

Hornibrook, N.de B., Brazier, R.C., Strong, C.P., 1989: Manual of New Zealand Permian to Pleistocene Foraminiferal Biostratigraphy. New Zealand Geological Survey Paleontological Bulletin, no. 56. 175p

Hudson, D.S., 1996: *Provenance of Miocene to Recent sands and sandstones of the Taranaki – Wanganui region, New Zealand.* M.Sc. thesis, University of Waikato, New Zealand.

Hunt, D., Tucker, M.E., 1992: Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall. *Sedimentary Geology, 81:* 1-9.

Hunt, D., Tucker, M.E., 1993: Sequence stratigraphy of carbonate shelves with an example from the mid-Cretaceous (Urgonian) of southeast France. *In* Posamentier, H.W., Summerhayes, C.P, Haq, B.U., and Allen, G.P. (eds) Sequence Stratigraphy and Facies Associations. *International Association of Sedimentologists Special Publication no. 18:* 307-341

Hunt, D., Tucker, M.E., 1995: Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall - reply. *Sedimentary Geology, 95:* 147-1609.

Jervey, M.T., 1988: Quantitative geological modeling of siliciclastic rock sequences and their seismic expression. Wilgus, B.S., Hastings, C.G. St Kendall, Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (eds) Sea-level Changes: an Integrated Approach. *Special Publication, Society of Economic Paleontologists and Mineralogists, Tulsa, 42:* 47-69.

Jordan, D.W., Schultz, D.J., Cherng, J.A., 1994: Facies architecture and reservoir quality of Miocene Mt Messenger deep-water deposits, Taranaki Peninsula, New Zealand. *In* Weimer, P., Bouma, A.H., Perkins, B. (eds) Submarine fans and turbidite systems: sequence stratigraphy, reservoir architecture and production characteristics, Gulf of Mexico and international: 15 1-166. *Gulf Coast Section, Society of Economic Paleontologists and Mineralogists Foundation, Houston*

Kamp, P.J.J., 1991: An investigation of the thermal and tectonic history of Pluto-1 hydrocarbon exploration well, Taranaki Basin, New Zealand using fission track analysis. Fission Track Research Group Commercial Report #3. Geochronology Research Unit, University of Waikato, New Zealand. 278p.

Kamp, P.J.J., Green, P.F., White, S.H., 1989: Fission track analysis reveals character of collisional tectonics in New Zealand. *Tectonics 8:* 169-195.

Kamp, P.J.J., McIntyre, A.P., 1998: The stratigraphic architecture of Late Pliocene (2.8-2.4 Ma) asymmetrical shelf sequences, western Wanganui Basin, New Zealand. *Sedimentary Geology, 122:* 53-67.

Kamp, P.J.J., Naish, T., 1998: Forward modeling of the sequences stratigraphic architecture of shelf cyclothems: application to Late Pliocene sequences, Wanganui Basin (New Zealand). *Sedimentary Geology, 116:* 57-80.

Kamp, P.J.J., Tippett, J.M., 1993: Dynamics of Pacific Plate crust in the South Island (New Zealand) zone of oblique continent-continent convergence. *Journal of Geophysical Research, vol. 98, no. B9:* 16105 – 16118.

Kamp, P.J.J., Vonk, A.J., Bland, K.J., Griffin, A.G., Hayton, S., Hendy, A.J.W., McIntyre, A.P., Nelson, C.S., Naish, T., 2002: Megasequence architecture of Taranaki, Wanganui, and King Country Basins and Neogene progradation of two continental margin wedges across western New Zealand. *In* 2002 New Zealand Petroleum Conference proceedings: 464-481. Ministry of Economic Development, Wellington.

Kamp, P.J.J., Vonk, A.J., Bland, K.J., Griffin, A.G., Hayton, S., Hendy, A.J.W., McIntyre, A.P., Nelson, C.S., Naish, T., in prep.: Geology of Wanganui Basin and its petroleum systems. The University of Waikato, New Zealand.

Kearey, P., Brooks, M., 1991: *An introduction to geophysical exploration* (2nd Edition). Blackwell Scientific Publications, Oxford. 254 p.

Kennett, J.P., Srinivasan, M.S., 1983: *Neogene Planktonic Foraminifera: A phylogenetic atlas.* Hutchinson Ross Publishing Company, New York. 265 p

King, P.R., 1988a: Well Summary Sheets, onshore Taranaki. New Zealand Geological Survey report G125.

King, P.R., 1988b: Well Summary Sheets, offshore Taranaki. New Zealand Geological Survey report G127.

King, P.R., 1990: Polyphase evolution of the Taranaki basin, New Zealand: changes in sedimentary and structural style. *In* 1989 New Zealand Oil Exploration Conference proceedings: 134-150. Ministry of Commerce, Wellington.

King, P.R., 1992: Outcropping Miocene sandstones, North-east Taranaki Basin: A look at sequences with subsurface reservoir potential. *Petroleum Exploration in New Zealand News, 32:* 27-32.

King, P.R., 2000a: New Zealand's changing configuration in the last 100 million years: plate tectonics, basin development, and depositional setting. *In* 2000 New Zealand Petroleum Conference proceedings: 131-145. Ministry of Commerce, Wellington

King. P.R., 2000b: Tectonic reconstructions of New Zealand: 40 Ma to the Present. *New Zealand Journal of Geology and Geophysics, 43:* 611-638.

King, P.R., Browne, G.H., Slatt, R.M., 1994: Sequence architecture of exposed late Miocene basin floor fan and channel-levee complexes (Mount Messenger Formation), Taranaki Basin, New Zealand. *In* Weimer, P., Bouma, A.H., Perkins, G.F. (eds) Submarine Fans and Turbidite Systems: sequence stratigraphy, reservoir architecture and production characteristics Gulf of Mexico and international: 172-192. *Gulf Coast Section, Society of Economic Paleontologists and Mineralogists Foundation, Houston.*

King, P.R., Scott, G.H., Robinson, P.H., 1993: Description, correlation and depositional history of Miocene sediments outcropping along North Taranaki coast. *Institute of Geological & Nuclear Sciences Monograph 5.* 199 p. Institute of Geological and Nuclear Sciences Ltd., Lower Hutt.

King, P.R., Thrasher, G.P., 1992: Post-Eocene development of the Taranaki Basin, New Zealand: Convergent overprint of a passive margin. *In* Watkins, J.S., Zhiqiang, F., and McMillan, K.J. (eds) Geology and Geophysics of Continental Margins. *American Association of Petroleum Geologists Memoir 53:* 93-118.

King, P.R., Thrasher, G.P., 1996: Cretaceous-Cenozoic geology and petroleum systems of the Taranaki Basin, New Zealand. *Institute of Geological and Nuclear Sciences monograph 13.* 243 p. 6 enclosures. Institute of Geological & Nuclear Sciences Ltd., Lower Hutt.

Kirschner, R.H., Bouma, A.H., 2000: Characteristics of a distributary channel-levee-overbank system, Tanqua Karoo. *In* Bouma, A.H., Stone, C.G. (eds) Fine-grained turbidite systems. *AAPG Memoir 72/SEPM Special Publication no. 68:* 233-244.

Knox, G.J., 1982: Taranaki Basin, structural style and tectonic setting. *New Zealand Journal of Geology and Geophysics, 25:* 51-60.

Kolla, V., Posamentier, H.W., Eichenseer, H., 1995: Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall – discussion. *Sedimentary Geology 95:* 139-145.

Kovach, W.L., 1999: *MSVP – A multivariate statistical package for Windows, ver. 3.1.* Kovach Computing Services, Pentraeth, Wales, U.K. 133 p.

Lawrence, D.T., 1993: Evaluation of eustasy, subsidence, and sediment input as controls on depositional sequence geometries and the synchroneity of sequence boundaries. *In* Weimer, P., and Posamentier, H. (eds) Siliciclastic sequence stratigraphy: recent developments and applications. *American Association of Petroleum Geologists memoir 58.* 337-367.

Leeder, M., Raiswell, R., Al-Biatty, H., McMahon, A., Hardman, M., 1990: Carboniferous stratigraphy, sedimentation and correlation of well 48/3-3 in the southern North Sea Basin: integrated use of palynology, natural gamma/sonic logs and carbon/sulphur geochemistry. *Journal of the Geological Society, London, 147:* 287-300.

Leslie, A.B., Spiro, B., Tucker, M.E., 1993: Geochemical and mineralogical variations in the upper Mercia Mudstone Group (Late Triassic), southwest Britain: correlation of outcrop sequences with borehole geophysical logs. *Journal of the Geological Society, London, 150:* 67-75.

Lewis, D.W., McConchie, D., 1994: Analytical sedimentology. Chapman and Hall, New York. 197 p.

Lewis, K.B., 1979: Foraminifera on the Continental shelf and slope off Southern Hawkes Bay, New Zealand. *New Zealand Oceanographic Institute Memoir no. 84.* 45p.

Lindseth, R.O., Beraldo, V.L., 1985: A Late Cretaceous submarine canyon in Brazil. *In:* Berg, O.R., and Woolverton, D.G. (eds) Seismic Stratigraphy II: An integrated approach to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 39:* 169-180.

Loeblich, A.R., Jr., Tappan, H., 1987: *Foraminiferal genera and their classification.* 2 vols, Van Nostrand Reinhold, Melbourne, Australia. 1182 p.

Lohmann, G.P., 1995: A model for variation in the chemistry of planktonic foraminifera due to secondary calcification and selective dissolution. *Paleoceanography, vol. 10, no. 3:* 445-457.

Loutit, T.S., Hardenbol, J., Vail, P.R., Baum, G.R., 1988: Condensed sections: the key to age determination and correlation of continental margin sequences. *In* Wilgus, C. K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (eds) Sea level changes: an integrated approach. *Society of Economic Paleontologists and Mineralogists Special Publication no. 42:* 183-213.

Lovell, M.A., Harvey, P.K., Jackson, P.D., Brewer, T.S., Williamson, G., Williams, C.G., 1998: Interpretation of core and log data – integration or calibration? *In* Harvey, P.K., & Lovell, M.A. (eds) Core-Log Integration, Geological Society, London. *Special Publications 136:* 39-51.

May, J.A., Warme, J.E., Slater, R.A., 1983: Role of submarine canyons on shelfbreak erosion and sedimentation: modern and ancient examples. *In* Stanley, D.J., Moore, G.T., (eds) The shelfbreak: critical interface on continental margins. *Society of Economic Paleontologists and Mineralogists Special Publication no. 33*: 315-332.

McAlpine, A., 2000: Basin modelling of oil plays northwest of Maui: results, constraints and calibrations. *In* 2000 New Zealand Petroleum Conference proceedings: 59-67. Ministry of Commerce, Wellington.

McCartney, M.S., 1982: The subtropical recirculation of Mode Waters. *Journal of Marine Research, 40 (Supplement):* 427-464.

McIntyre, A.P., 2001: *Geology of Mangapanian (Late Pliocene) strata, Wanganui Basin: Lithostratigraphy, paleontology and sequence stratigraphy.* Unpublished PhD thesis, The University of Waikato.

McQuillin, R., Bacon, M., Barclay, W., 1984: *An introduction to seismic interpretation. Reflection seismics in petroleum exploration.* Graham and Trotman Limited, London. 287 p.

Metzger, J.M., Flemings, P.B., Christie-Blick, N., Mountain, G.S., Austin Jr, J.A., Hesselbo, S.P., 2000: Late Miocene to Pleistocene sequences at the New Jersey outer continental shelf (ODP leg 174A, sites 1071 and 1072). *Sedimentary Geology, 134:* 149-180.

Miall, A.D., 1997: *The geology of stratigraphic sequences.* Springer-Verlag, Heidelberg, Germany. 433 p.

Mills, C., 1989: Gravity expression of the Patea-Tongaporutu High and subsequent model for the Taranaki Basin margin. *In* 1988 New Zealand Oil Exploration Conference proceedings: 191-200. Ministry of Commerce, Wellington.

Milton, N.J., Bertram, G.B., 1995: Tectonic controls on systems tract development: implications for hydrocarbon exploration. *In* Hesselbo, S.P., Parkinson, D.N. (eds) Sequence Stratigraphy and its Application to the British Stratigraphic Record. *Special Publication, Geological Society of London.*

Mitchum, R.M., 1985: Seismic stratigraphic expression of submarine fans. *In* Berg, O.R., and Woolverton, D.G. (eds) Seismic Stratigraphy II: An integrated approach to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 39:* 117-136.

Mitchum, R.M. Jr., Vail, P.R., Sangree, J.B., 1977a: Seismic stratigraphy and global changes in seal level, Part 6: Stratigraphic interpretation of seismic reflection patterns in depositional sequences. *In* Payton, C.E. (ed) Seismic stratigraphy – applications to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 26:* 117-133.

Mitchum, R.M. Jr., Vail, P.R., Thompson, S. III, 1977b: Seismic stratigraphy and global changes in seal level, Part 2: The depositional sequence as a basic unit for stratigraphic analysis. *In* Payton, C.E. (ed) Seismic stratigraphy – applications to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 26:* 53-62.

Mitchum, R.M., Sangree, J.B., Vail, P.R., Wornardt, W.W., 1993: Recognizing sequences and systems tracts from well logs, seismic data, and biostratigraphy: Examples from the Late Cenozoic of the Gulf of Mexico. *In* Weimer, P., and Posamentier, H. (eds) Siliciclastic sequence stratigraphy: recent development and applications. *American Association of Petroleum Geologists memoir 59.* 163-197.

Morgans, H.E.G., Scott, G.H., Beu, A.G., Graham, I.J., Mumme, T.C., St George, W., Strong, C.P., 1997: New Zealand Cenozoic Timescale (version 11/96). Institute of Geological & Nuclear Sciences science report 96/38. 12p.

Mortimer, N., Tulloch, A.J., Ireland, T.R., 1997: Basement geology of Taranaki and Wanganui Basins, New Zealand. *New Zealand Journal of Geology and Geophysics, 40:* 223-236.

Murray, D., de Bock, J.F., 1996: Awatea-1 well completion Report. PEP 38457. Ministry of commerce New Zealand Unpublished Petroleum Report 2262.

Murray, J.W., 1973: *Distribution and ecology of living benthic foraminiferids.* Heinemann Education Books Limited, Great Britain. 288 p.

Murray, J.W., 1976: A method of determining proximity of marginal seas to an ocean. *Marine Geology, 22:* 103-119.

Murray, J.W., 1991a: Ecology and distribution of benthic foraminifera. *In* Lee, J.J., and Anderson, O.R. (eds) *Biology of Foraminifera.* Academic Press, London. 221-253.

Murray, J.W, 1991b: Ecology and distribution of planktonic foraminifera. *In* Lee, J.J., and Anderson, O.R. (eds) *Biology of Foraminifera.* Academic Press, London. 254-284.

Murray, J.W., 1992: Ecology and distribution of benthic foraminifera: A review. *In* Takayanagi, T., and Saito, T. (eds) *Studies in Benthic Foraminifera: Proceedings of the fourth international symposium on benthic foraminifera, Sendai, 1990.* Tokai University Press, Japan. 33-41.

Murray, J.W., 1995: Microfossil indicators of ocean water masses, circulation and climate. In Bosence, D.W.J, Allison, P.A. (eds) Marine Palaeoenvironmental analysis from fossils. *Geological Society Special Publication no. 83*: 245-264.

Murray, J.W., 2001: The niche of benthic foraminifera, critical thresholds and proxies. *Marine Micropaleontology, 41*: 1-7.

Mutti, E., Normark, W.R., 1987: Comparing examples of modern and ancient turbidite systems: problems and concepts. *In* Leggett, J.K., Zuffa, G.G.. (eds) *Marine clastic sedimentology: concepts and case studies*: 1-38. Graham & Trotman, London.

Naish, T., Kamp, P.J.J., 1995: Plio-Pleistocene marine cyclothems, Wanganui Basin, New Zealand. *Sedimentary Geology, 110:* 237-255.

Naish, T., Kamp, P.J.J., 1997a: Foraminiferal depth palaeoecology of Late Pliocene shelf sequences and systems tracts, Wanganui Basin, New Zealand. *Sedimentary Geology, 110*: 237-255.

Naish, T., Kamp, P.J.J., 1997b: Sequence stratigraphy of 6th order (41 k.y.) Pliocene-Pleistocene cyclothems, Wanganui Basin, New Zealand: a case for the regressive systems tract. *Geological Society of America Bulletin, 109:* 978-999.

Nathan, S., 1997: New Zealand stage names. Institute of Geological and Nuclear Sciences Newsletter 6(4): 3.

Nelson, C.S., Cooke, P.J., 2001: History of oceanic front development in the New Zealand sector of the Southern Ocean during the Cenozoic - a synthesis. *New Zealand Journal of Geology and Geophysics, 44:* 535-553.

Nelson, C.S., Kamp, P.J.J., Young, H.R., 1994: Sedimentology and petrography of mass-emplaced limestone (Orahiri Limestone) on a late Oligocene shelf, western North Island, and tectonic implications for eastern margin development of Taranaki Basin. *New Zealand Journal of Geology and Geophysics, 37:* 269-285.

Nodder, S.D., 1987: *The mid Miocene geology of the Waikawau region, North Taranaki, New Zealand.* Unpublished MSc. thesis, University of Waikato.

Nodder, S.D., 1993: Neotectonics of the offshore Cape Egmont Fault Zone, Taranaki Basin, New Zealand. *New Zealand Journal of Geology and Geophysics, 36*: 167-184.

Nodder, S.D., Nelson, C.S., Kamp, P.J.J., 1990a: Mass emplaced siliciclastic-volcaniclastic-carbonate sediments in middle Miocene shelf-to-slope environments at Waikawau, northern Taranaki, and some implications for Taranaki Basin development. *New Zealand Journal of Geology and Geophysics, 33:* 599-615.

Nodder, S.D., Nelson, C.S., Kamp, P.J.J., 1990b: Middle Miocene formational stratigraphy (Mokau-Mohakatino Groups) at Waikawau, northeastern Taranaki Basin margin, New Zealand. *New Zealand Journal of Geology and Geophysics, 33:* 585-598.

Nummedal, D., Riley, G.W., Templet, P.L., 1993: High-resolution sequence architecture: a chronostratigraphic model based on equilibrium profile studies. *In* Posamentier, H.W., Summerhayes, C.P., Haq, B.U., Allen, G.P. (eds) Sequence stratigraphy and facies associations. *Special Publication of the International Association of Sedimentologists, no. 18:* 55-68.

NZ Oil & Gas Services Ltd. (NZOG), 1990: Kahawai-1 well completion Report. PPL 38451. Ministry of Commerce New Zealand Unpublished Petroleum Report 1878.

NZ Petroleum Co Ltd., 1943a: Uruti-2. Ministry of Commerce New Zealand Unpublished Petroleum Report 217.

NZ Petroleum Co Ltd., 1943b: Uruti-1. Ministry of Commerce New Zealand Unpublished Petroleum Report 218.

Ogilvie, M.J., 1993: *The Pliocene-Pleistocene seismic stratigraphy of part of the offshore South Taranaki and South Wanganui Basins.* Unpublished M.Sc. thesis, Victoria University of Wellington.

Ottens, J.J., Nederbragt, A.J., 1992: Planktic foraminiferal diversity as indicator of ocean environments. *Marine Micropaleontology, 19:* 12-28.

Pekar, S.F., Miller, K.G., Kominz, M.A., 2000: Reconstructing the stratal geometry of latest Eocene to Oligocene sequences in New Jersey: resolving a patchwork distribution into a clear pattern of progradation. *Sedimentary Geology, 134:* 93-109.

Petrocorp Petroleum Corporation of New Zealand Exploration Ltd., 1995: Offshore 2D seismic survey 1995/1996 P95 lines, PPL 38455, PPL 38456, PPL 38457, PPL 38458, PPL 38459. Unpublished Petroleum report 2261. Ministry of Commerce, Wellington.

Petroleum Engineering and Exploration Departments, Shell BP Todd Oil Services Ltd., 1981: Well resume Tangaroa-1 PPL 38038 offshore. Ministry of Commerce, New Zealand. Unpublished Petroleum Report 793.

Pickering, K.T., Clark, J.D., Smith, R.D.A., Hiscott, R.N., Ricci Lucchi, F., Kenyon, N.H., 1995: Architectural element analysis of turbidite systems, and selected topical problems for sand-prone deep-water systems. *In* Pickering, K.T., Hiscott, R.N., Kenyon, N.H., Ricci Lucchi, F., Smith, R.D.A. (eds) *Atlas of deep-water environments: architectural style in turbidite systems.* Chapman and Hall, London. 1-10.

Pilaar, W.F.H., Wakefield, L.L., 1978: Structural and stratigraphic evolution of the Taranaki Basin, offshore North Island, New Zealand. *APEA Journal, vol. 18, Part 1:* 93-101.

Pirmez, C, Pratson, L.F., Steckler, M.S., 1998: Clinoform development by advection-diffusion of suspended sediment: modeling and comparison to natural systems. *Journal of Geophysical Research, vol. 103, no. B10:* 24141-24157.

Posamentier, H.W., Allen, G.P., 1993: Variability of the sequence stratigraphic model: effects of local basin factors. *Sedimentary Geology, 86:* 91-109.

Posamentier, H.W., Allen, G.P., James, D.P., Tesson, M., 1992: Forced regressions in a sequence stratigraphic framework: concepts, examples, and exploration significance. *The American Association of Petroleum Geologists Bulletin, vol. 76, no. 11:* 1687-1709.

Posamentier, H.W., Erskine, R.D., Mitchum, R.M., Jr., 1991: Models of submarine-fan deposition within a sequence stratigraphic framework. *In* Weimer, P., and Lin, M.H. (eds) *Seismic facies and sedimentary processes of submarine fans and turbidite systems:* 127-136. Springer-Verlag New York, Inc.

Posamentier, H.W., James, D.P., 1993: An overview of sequence-stratigraphic concepts: uses and abuses. *In* Posamentier, H.W., Summerhayes, C.P., Haq, B.U., Allen, G.P. (eds) Sequence stratigraphy and facies associations. *Special Publication of the International Association of Sedimentologists, no. 18:* 3-18.

Posamentier, H.W., Jervery, M.T., Vail, P.R., 1988a: Eustatic controls on clastic deposition I – conceptual framework. *Society of Economic Paleontologists and Mineralogists, special publication 42:* 109-124

Posamentier, H.W., Jervery, M.T, Vail, P.R., 1988b: Eustatic controls on clastic deposition I – conceptual framework: Stratigraphic interpretation of seismic reflection patterns in depositional sequences. *In* Wilgus, C. K., Hastings, B.S., Kendall, C.G.St.C, Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (eds) Sea level changes: an integrated approach. *Society of Economic Paleontologists and Mineralogists Special Publication no. 42:* 109-124

Posamentier, H.W., Vail, P.R., 1988: Eustatic controls on clastic deposition II – sequence and systems tract models. *In* Wilgus, C. K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (eds) Sea level changes: an integrated approach. *Society of Economic Paleontologists and Mineralogists Special Publication no. 42:* 125-154

Prather, B.E., Booth, J.R., Steffens, G.S., Craig, P.A., 1998: Classification, lithologic calibration, and stratigraphic succession of seismic facies of intraslope basins, deep-water Gulf of Mexico. *American Association of Petroleum Geologists Bulletin, vol. 82, no. 5A:* 701-728.

Pratson, LF., 2001: A perspective on what is known and not known abut seafloor instability in the context of continental margin evolution. *Marine and Petroleum Geology, 18:* 499-501.

Proctor, R., Carter, L., 1989: Tidal and sedimentary response to the Late Quaternary closure and opening of Cook Strait, New Zealand: Results from numerical modelling. *Paleoceanography, vol. 4, no. 2:* 167-180.

Pulham, A.J., 1993: Variations in slope deposition, Pliocene-Pleistocene, offshore Louisiana, Northeast Gulf of Mexico. *In* Weimer, P., Posamentier, H.W. (eds) *Siliciclastic sequence stratigraphy: recent developments and applications*: 199-233.

Rankin, J., Barbaresig, G.G., 1988: Final well report Tua Tua-1. PPL 38087. Ministry of Commerce New Zealand Unpublished Petroleum report 1389.

Reading, H.G., Richards, M., 1994: Turbidite systems in deep-water basin margins classified by grain size and feeder system. *American Association of Petroleum Geologists Bulletin vol. 78, no. 5*: 792-822.

Reed, J.D., 1992: Exploration geochemistry of the Taranaki Basin with emphasis on Kora. *In* 1991 New Zealand Oil Exploration Conference Proceedings: 364-372. Ministry of Commerce, Wellington.

Reynolds, A.D., 1994: Sequence stratigraphy from core and wireline log data: the Viking Formation, Albian, south central Alberta, Canada. *Marine and Petroleum Geology, vol. 11, no. 3:* 258-282.

Rider, M.H., 1986: *The geological interpretation of well logs.* Blackie and Sons Ltd., Glasgow. 168 p.

Rider, M.H., 1990: Gamma-ray log shape used as a facies indicator: critical analysis of an oversimplified methodology. *In* Hurst, A., Lovell, M.A. & Morton, A.C. (eds) Geological Applications of Wireline Logs. *Geological Society Special Publication No. 48:* 27-37.

Robert, C., Stein, R., Acquaviva, M., 1986: Cenozoic evolution and significance of clay associations in the New Zealand region of the South Pacific, Deep Sea Drilling Project, Leg 90. *In* Blakeslee, J.H. (ed) *Initial Reports of the Deep Sea Drilling Project, Volume XC, Part 2:* 1225-1235.

Robinson, P.H., Morris, B.D., Scott, G.H., 1987: Lithologic log and micropaleontology of Manutahi-1 core, onshore south Taranaki, New Zealand. New Zealand Geological Survey report PAL 121.

Salimullah, A.R.M., Stow, D.A.V., 1992: Wireline log signatures of resedimented volcaniclastic facies, ODP Leg 129, West Central Pacific. *In* Hurst, A., Griffiths, C.M., & Worthington, P.F. (eds) Geological Applications of Wireline Logs II. *Geological Society Special Publication No. 65:* 87-97.

Sangree, J.B., Widmier, J.M., 1977: Seismic stratigraphy and global changes in seal level, Part 9: Seismic interpretation of clastic depositional facies. *In* Payton, C.E. (ed) Seismic stratigraphy – applications to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 26:* 165-184.

Schlager, W., 1993: Accommodation and supply – a dual control on stratigraphic sequences. *Sedimentary Geology, 86:* 111-136.

Schnitker, D., 1994: Deep-sea benthic foraminifers: food and bottom water masses. *In* Zahn, R., Pederson, T.F., Kaminski, M.A., Labeyne, L. (eds) Carbon cycling in the Global Ocean: constraints on the ocean's role in global change. *NATO ASI Series, 117*: 539-554.

Schröder, C.J., Scott, D.B., Medioli, F.S., 1987: Can smaller benthic foraminifera be ignored in paleoenvironmental analyses? *Journal of Foraminifera Research, vol. 17, no. 2:* 101-105.

Scott, G.H., 1989: Report on Upper Miocene-Pliocene well biostratigraphy – Taimana-1, Kaimiro-1, New Plymouth-1 Te Kiri-1, Cape Egmont-1. Ministry of Commerce New Zealand Unpublished Petroleum Report 1486.

Scott, G.H., 2001: Local Stages and bioevents in time scale research. Unpublished memo. Institute of Geological and Nuclear Sciences, Lower Hutt.

Scott, G.H., Bishop, S., Burt, B.J., 1990: Guide to some Neogene Globorotalids (Foraminiferida) from New Zealand. New Zealand Paleontological Bulletin no. 61. New Zealand Geological Survey, Lower Hutt. 135 p.

Scott, G.H., Crundwell, M.P., 1994: Pliocene foraminiferal environments in Hawke Bay-1 well. Institute of Geological & Nuclear Sciences science report 94/19. 27p.

Scott, G.H., King, P.R., Crundwell, M.P., in prep.: Relation between seismic and foraminiferal facies in a Pliocene progradational shelf margin complex, Taranaki Basin, New Zealand. Institute of Geological and Nuclear Sciences, Lower Hutt.

Sen Gupta, B.K., Machain-Castillo, M.L., 1993: Benthic foraminifera in oxygen-poor habitats. *Marine Micropaleontology, 20*: 183-201.

Shanmugam, G., 2000: 50 years of the turbidite paradigm (1950s-1990s): deep-water processes and facies models – a critical perspective. *Marine and Petroleum Geology, 17*: 285-342.

Shanmugam, G., Bloch, R.B., Mitchell, S.M., Beamish, G.W.J., Hodgkinson, R.J., Damuth, J.E., Straume, T., Syvertsen, S.E., Shields, K.E., 1995: Basin-floor fans in the North Sea: sequence stratigraphic modes vs. sedimentary facies. *The American Association of Petroleum Geologists Bulletin, vol. 79, no. 4:* 477-512.

Shanmugam, G., Moiola, R.J., 1991: Types of submarine fan lobes: models and implications. *The American Association of Petroleum Geologists Bulletin vol. 75, no. 1:* 156-179.

Shell BP Todd Oil Services Ltd., 1975: Well resume Turi-1. Ministry of Economic Development, New Zealand. Unpublished Petroleum Report 659.

Shell BP Todd Oil Services Ltd., 1976: Well resume Tane-1 (offshore). Ministry of Commerce, New Zealand. Unpublished Petroleum Report 698.

Shell BP Todd Oil Services Ltd., 1981: Well resume Tangaroa-1, PPL 38038, offshore Taranaki. Ministry of Commerce, New Zealand Unpublished Petroleum Report 793.

Shell BP Todd Oil Services Ltd., 1982: Well resume, Wainui-1 PPL 38049. Ministry of Commerce, New Zealand Unpublished Petroleum Report 869.

Shell BP Todd Oil Services Ltd., 1984: Completion Report, Ariki-1, PPL 38048. Ministry of Commerce, New Zealand. Unpublished Petroleum Report 1038.

Shell Oil Company, 1987: Seismic sequences as applied to basin-fill analysis: Taranaki Basin, New Zealand. *In* Bally, A.W. (ed) Atlas of Seismic Stratigraphy, volume 1. *American Association of Petroleum Geologists Studies in Geology, no. 27:* 53-71.

Shell Todd Oil Services Ltd., 1991: Well resume Pluto-1, PPL 38098 & PPL 38453, offshore north Taranaki, NZ. Ministry of Commerce New Zealand Unpublished Petroleum Report 1766.

Sheriff, R.E., 1985: Aspects of seismic resolution. *In* Berg, O.R., and Woolverton, D.G. (eds), Seismic Stratigraphy II: An integrated approach to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 39:* 1-10.

Sheriff, R.E., 1989: *Geophysical Methods*. Prentice Hall, New Jersey. 605 p.

Sherwood. A., 1997: New Zealand stage names. Institute of Geological and Nuclear Sciences Newsletter 6(4): 3.

Sircombe, K.N., 1993: Analysis of the South Westland Basin, New Zealand. Unpublished MSc. thesis, University of Waikato.

Slatt, R.M., Jordan, D.W., D'Agostino, A.E., Gillespie, R.H., 1992. Outcrop gamma ray logging to improve understanding of subsurface well log correlations. *In* Hurst, A., Griffiths, C.M., & Worthington, P.F. (eds) Geological Applications of Wireline Logs II. *Geological Society Special Publication No. 65:* 3-19.

Smale, D., 1992: Provenance of sediments in Taranaki Basin – an assessment from heavy minerals. *In* 1991 New Zealand Oil Exploration Conference proceedings: 245-254. Ministry of Commerce, Wellington.

Smale, D., 1996: Petrographic summaries of Taranaki petroleum reports. Institute of Geological & Nuclear Sciences science report 96/1. 88p.

Soenander, H.B., 1991: *Seismic stratigraphy of the Giant Foresets Formation, north Taranaki Basin.* Unpublished M.Sc. thesis, Victoria University of Wellington, New Zealand.

Soenander, H.B., 1992: Seismic stratigraphy of the Giant Foreset Formation, offshore North Taranaki–Western Platform. *In* 1991 New Zealand Oil Exploration Conference proceedings: 207-233. Ministry of Commerce, Wellington.

Stagpoole, V., 1997: *A geophysical study of the northern Taranaki Basin, New Zealand.* Unpublished PhD thesis, Victoria University of Wellington, New Zealand.

Stagpoole, V., 2000: Fluid flow and pressure development associated with foreset progradation. (presented at petroleum workshop, New Plymouth, 2000). Institute of Geological and Nuclear Sciences Limited, Lower Hutt.

Stanton, B.R., 1973: Circulation along the eastern boundary of the Tasman Sea. *In* Fraser, R. (ed) Oceanography of the South Pacific 1972. Wellington, New Zealand National Commission for UNESCO. 141-147.

Steckler, M.S., Reynolds, D.J., Coakley, B.J., Swift, B.A., Jarrard, R., 1993: Modelling passive margin sequence stratigraphy. *In* Posamentier, H.W., Summerhayes, C.P., Haq, B.U., Allen, G.P. (eds) Sequence stratigraphy and facies associations. *Special Publication of the International Association of Sedimentologists, no. 18:* 19-41.

Stelting, C.H., Bouma, A.H., Stone, C.G., 2000: Fine-grained turbidite systems: Overview. *In* Bouma, A.H., Stone, C.G. (eds) Fine-grained turbidite systems. *American Association of Petroleum Geologists Memoir 72/SEPM Special Publication no. 68:* 1-8.

Stern, T.A., Davey, F.J., 1990: Deep seismic expression of a foreland basin: Taranaki Basin, New Zealand. *Geology, 18:* 979-982.

Stow, D.A.V., Mayall, M., 2000: Deep-water sedimentary systems: New models for the 21st century. *Marine and Petroleum Geology, 17*: 125-135.

Strong, C.P., Waghorn, D.B., Cooper, R.A., Crampton, J.S, Morgans, H.E.G., Raine, J.I., 1996: biostratigraphy of Awatea-1 offshore petroleum exploration drillhole, North Taranaki Basin. Institute of Geological and Nuclear Sciences Client Report 53652A.10. Institute of Geological and Nuclear Sciences, Lower Hutt, New Zealand.

Sykes, R., Suggate, R.P., King, P.R., 1992: Timing and depth of maturation in southern Taranaki Basin form reflectance and rank(s). *In* 1991 New Zealand Oil Exploration Conference proceedings: 373-289. Ministry of Commerce, Wellington.

Ter Braak, C.J.F., 1986: Canonical correspondence analysis: a new eigenvector technique for multivariate direct gradient analysis. *Ecology, vol. 67, no.* 5: 1167-1179.

Thrasher, G.P., 1990: Tectonics of the Taranaki Rift. *In* 1989 New Zealand Oil Exploration Conference proceedings: 124-133. Ministry of Commerce, Wellington.

Thrasher, G.P., 1992: Late Cretaceous source rocks of Taranaki Basin. *In* 1991 New Zealand Oil Exploration Conference proceedings: 147-154. Ministry of Commerce, Wellington.

Thrasher, G.P., Cahill, J.P., 1990: Subsurface maps of the Taranaki Basin region 1:50 000, New Zealand. New Zealand Geological Survey Report G-142.

Thrasher, G.P., Leitner, B., Hart, A.W., 2002: Petroleum system of the Northern Taranaki Graben. *In* 2002 New Zealand Petroleum Conference Proceedings: 1-6. Ministry of Economic Development, Wellington.

Tippett, M.J., Kamp, P.J.J., 1993: Fission track analysis of the late Cenozoic vertical kinematics of continental Pacific crust, South Island, New Zealand. *Journal of Geophysical Research, vol. 98, no. B9:* 16119 – 16148.

Todd Oil Services Ltd 1984 (ariki-1)

Vail, P.R., Mitchum, R.M. Jr., Thompson, S. III, 1977a: Seismic stratigraphy and global changes in seal level, Part 3: Relative changes of sea level from coastal onlap. *In* Payton, C.E. (ed) Seismic stratigraphy – applications to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 26:* 63-81.

Vail, P.R., Mitchum, R.M. Jr., Thompson, S. III, 1977b: Seismic stratigraphy and global changes in seal level, Part 4: Global cycles of relative changes of sea level. *In* Payton, C.E. (ed) Seismic stratigraphy – applications to hydrocarbon exploration. *American Association of Petroleum Geologists memoir 26:* 83-97.

Vail, P.R., 1987: Seismic stratigraphy interpretation using sequence stratigraphy Part I: Seismic stratigraphy interpretation procedure. *In* Bally, A.W. (ed) AAPG Atlas of Seismic Stratigraphy, volume 1. *American Association of Petroleum Geologists studies in geology no. 27:* 1-10. American Association of Petroleum Geologists, Tulsa.

Vail, P.R., Audemart, F., Bowman, S.A., Eisner, P.N., Perez-Cruz, G., 1991: The stratigraphic signatures of tectonics, eustasy and sedimentation - an overview. *In* Einsele, G., Ricken, W., Seilacher, A. (eds) *Cyclic stratigraphy.* Springer-Verlag, New York: 617-659.

Van der Zwaan, G.J., Duijnstee, I.A.P., den Dulk, M., Ernst, S.R., Jannik, N.T., Kouwenhoven, T.J., 1999: Benthic foraminifers: proxies or problems? A review of paleocological concepts. *Earth-Science Reviews, 46*: 213-236.

Van Morkhoven, F.P.C.M., Berggren, W.A., Edwards, A.S., 1986: Cenozoic cosmopolitan deep-water benthic foraminifera. *Bulletin Des Centers Des Recherches Exploration-Production Elf-Aquitaine Memoir 11.* 421 p

Vanney, J-R., Stanley, D.J., 1983: Shelfbreak physiography: an overview. *In* Stanley, D.J., Moore, G.T. (eds) The shelfbreak: critical interface on continental margins. *Society of Economic Paleontologists and Mineralogists Special Publication no. 33*: 1-24.

Van Wagoner, J.C., Mitchum, R.M., Jr., Campion, K.M., Rahmanian, V.D., 1990: Siliciclastic sequence stratigraphy in well logs, cores and outcrop: concepts for high resolution correlation of time and facies. *American Association of Petroleum Geologists Methods in Exploration Series, Tulsa, 7.* 55 p.

Van Wagoner, J.C., Posamentier, H.W., Mitchum, R.M., Jr., Vail, P.R., Sarg, J.F., Loutit, T.S., Hardenbol, J., 1988: An overview of the fundamentals of sequence stratigraphy and key definitions. *In* Wilgus, C. K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (eds) Sea level changes: an integrated approach. *Society of Economic Paleontologists and Mineralogists Special Publication no. 42:* 39-45

Vella, P., 1957: Studies in New Zealand Foraminifera. *New Zealand Geological Survey Paleontological Bulletin no. 28.*

Villamil, T., Arango, C., Weimer, P., Waterman, A., Rowan, M.G., Varnai, P., Pulham, A.J., Crews, J.R., 1998: Biostratigraphic techniques for analyzing benthic biofacies, stratigraphic condensation, and key surface identification, Pliocene and Pleistocene sediments, Northern Green Canyon and Ewing Bank (offshore Louisiana), Northern Gulf of Mexico. *American Association of Petroleum Geologists Bulletin, vol. 82, no. 5B, Part B:* 961-985.

Vonk, A.J., Kamp, P.J.J., Hendy, A.J.W., 2002: Outcrop to subcrop correlations of late Miocene-Pliocene strata, eastern Taranaki Peninsula. *In* 2002 New Zealand Petroleum Conference proceedings: 1-22. Ministry of Economic Development, Wellington.

Waghorn, D.B., Strong, C.P., Raine, J.I., Crampton, J.S., 1996: Biostratigraphic review of the Late Miocene and Pliocene of the Mangaa-1, Kahawai-1, Te Kumi-1, Tangaroa-1, and Kora-1 wells, offshore Taranaki Basin. Ministry of Commerce New Zealand Unpublished Petroleum Report 2417.

Walker, R.G., 1992: Facies, facies models, and modern stratigraphic concepts. *In* Walker, R.G. and James, N.P. (eds) Facies models: response to sea level change. *Geological Association of Canada.* 1-14.

Walker, R.G., 1992: Turbidites and submarine fans. *In* Walker, R.G. and James, N.P. (eds) Facies models: response to sea level change. *Geological Association of Canada.* 239-263.

Waltham, D., Taberner, C., Docherty, C., 2000: Error estimation in decompacted subsidence curves. *American Association of Petroleum Geologists Bulletin, vol. 84, no. 8:* 1087-1094.

Wehr, F.L., 1993: Effects of variation in subsidence and sediment supply on parasequence stacking patterns. *In* Weimer, P., Posamentier, H.W. (eds) Siliciclastic sequence stratigraphy: recent developments and applications. *American Association of Petroleum Geologists memoir 59*: 369-379.

Weimer, P., 1990: Sequence stratigraphy, facies geometries, and depositional history of the Mississippi Fan, Gulf of Mexico. *The America Association of Petroleum Geologists Bulletin, vol. 74, no. 4:* p425-453.

Weimer, P., Posamentier, H.W., 1993: Recent developments and applications siliciclastic sequence stratigraphy. *In* Weimer, P., Posamentier, H.W. (eds) Siliciclastic sequence stratigraphy: recent developments and applications. *American Association of Petroleum Geologists memoir 59*: 3-12.

Whittaker, A., 1998: Borehole data and geophysical log stratigraphy. *In* Doyle, P., Bennett, M.R. (eds) *Unlocking the stratigraphical record: Advances in modern stratigraphy.* John Wiley & Sons Ltd., England. 243-274.

Wilgus, C. K., Hastings, B.S., Kendall, C.G. St. C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (eds) 1988: Sea level changes: an integrated approach. *Society of Economic Paleontologists and Mineralogists Special Publication no. 42.* 407 p.

Williams, B.L., (ed) 1985: Ocean outfall handbook: A manual for the planning;, investigation, design an monitoring of ocean outfalls to comply with water quality management objectives. Water and Soil Miscellaneous Publication no. 76. 219 p.

Wilson, B.T., 1994: Sedimentology of the Miocene succession (coastal section), eastern Taranaki Basin margin: sequence stratigraphic interpretation. Unpublished MSc. thesis, University of Waikato.

Wonders, A.A.H., 1986: The biostratigraphy of the wells Tane-1 and Wainui-1, Taranaki Basin, offshore North Island, NZ. Ministry of Commerce New Zealand Unpublished Petroleum report 1184.

Worthington, P.F., 1990: Sediment cyclicity from well logs. *In* Hurst, A., Lovell, M.A. & Morton, A.C. (eds) Geological Applications of Wireline Logs. *Geological Society Special Publication No. 48:* 123-132.

Zimmerle, W., 1995: *Petroleum Sedimentology.* Kluwer Academic Publishers, Germany. 413p.