CLAY PETROLOGY
OF MESOZOIC TO RECENT SEDIMENTS
OF CENTRAL WESTERN NORTH ISLAND,
NEW ZEALAND.

BY

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Thesis for the degree of
Doctor of Philosophy
in
Earth Sciences.

University of Waikato
April, 1978

ABSTRACT

Clay mineral, bulk mineral and chemical data for Mesozoic to Recent onshore sediments and Recent western shelf surficial sediments from central Western North Island are presented with the aim of determining the significance of areal and stratigraphic variations in clay mineral distributions.

Sample analysis procedures using mainly X-ray diffractometry have been determined that are relatively simple, rapid and applicable to this wide variety of sediment types. Clay mineral analysis was carried out on the $<2\mu$ size fraction of all samples, and on the $<1\mu$, $2-4\mu$, $4-8\mu$ and $8-16\mu$ size fractions of selected samples to compensate for any mineralogic bias imposed through the selection of the traditional $<2\mu$ fraction. The predominance in the same samples of illite and chlorite in coarser fractions but of montmorillonite and kaolinite in finer fractions emphasises that the geologic interpretation of clay mineral distributions in sediments may be significantly influenced by the grain size fraction chosen for analysis. Clay mineral mounts prepared using dropper-on-glass slide and smear-on-glass slide techniques gave identical qualitative and similar quantitative results so that the former simpler technique was used.

Mesozoic to Recent sediments in Hamilton, Taranaki and Wanganui Districts contain varying amounts of kaolinite, illite, montmorillonite, chlorite and mixed-layer clay minerals. Clay mineral assemblages of mudstones, limestones or concretions, represent best those assemblages existing at the time of sediment deposition; clay minerals in sandstones are prone to diagenetic alteration because of their generally greater porosity.

The clay mineralogy of Mesozoic to Pleistocene sediments bears the imprint of a complex set of factors, including provenance, climate, tectonism, depositional environment and diagenesis. The illite, chlorite and mixed-layer clay mineral suite of Mesozoic sediments is consistent with derivation from an igneous/metamorphic provenance, partly modified by transformation during burial diagenesis. The abundant kaolinite in Upper Eocene Waikato Coal Measures is in part detrital from soils developed by prolonged chemical weathering of Mesozoic sediments, and in part diagenetic from clay mineral transformations under acid leaching conditions associated with carbonaceous sediments. The abundant crystalline montmorillonite in Oligocene sediments (Te Kuiti Group) was primarily neoformed in shallow marine basins characterised by extremely slow deposition and carbonate sedimentation. The abundant crystalline illite and chlorite characteristic of most Miocene to Lower Pleistocene sediments were mainly detrital from Mesozoic sediments under conditions of increasing topographic relief and more rapid physical erosion as a result of accelerating tectonic activity during this period. Montmorillonite in Lower Miocene sediments (Mahoenui and Mokau Groups) was largely detrital from uplifted Oligocene sediments. However, montmorillonite in Middle Miocene sediments (Mohakatino Group) was derived primarily from the alteration of contemporary volcanic products while much of the montmorillonite in Upper Miocene to Pliocene sediments (Mt Messenger to Tangahoe Formations) was detrital from Mohakatino lithologies. The increased abundance of montmorillonite in Plio-Pleistocene sediments results from the alteration of increasing quantities of volcanogenic material derived from contemporaneous volcanism in central and western North Island.

The clay mineralogy of western shelf sediments is dominated by illite and montmorillonite, with lesser amounts of chlorite and mixed-layer clays, and some kaolinite. On the basis of the distribution and origin of these clay minerals, five petrologic zones are distinguished: (1) Hamilton Shelf - Abundant montmorillonite, rare to common poorly crystalline illite, and lesser kaolinite, mixed-layer clays and poorly crystalline chlorite; primarily detrital from Oligocene mudstones, some montmorillonite neoformed from Quaternary volcanogenic material. (2) North Taranaki Shelf - Abundant crystalline illite, common mixed-layer clays and crystalline chlorite, minor montmorillonite; primarily detrital from Miocene mudstones. (3) Central Taranaki Shelf - Abundant crystalline illite, very common montmorillonite, common chlorite and mixed-layer clays, and some kaolinite. Neoformation of montmorillonite from Egmont-derived volcanogenic material contemporaneous with deposition of detrital illite, montmorillonite, chlorite and mixed-layer clays from northwestern South Island and perhaps North and South Taranaki Bight terrestrial provenances. (4) South Taranaki Shelf - Abundant crystalline illite, and common mixed-layer clays and crystalline chlorite; primarily detrital from northwestern South Island sources. (5) North Cook Strait Basin - Abundant crystalline illite, common mixed-layer clays and crystalline chlorite, and some montmorillonite; primarily detrital from northwestern South Island sources and, to a lesser extent, from Plio-Pleistocene mudstones in south Taranaki-Wanganui Districts. Clay mineral distributions on the western shelf are the product of a variety of interrelated physical, chemical and biochemical processes that have produced both detrital and authigenic sediments. Moreover, distribution patterns are further complicated by the existence of both modern and relict material.

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ACKNOWLEDGEMENTS

I am particularly indebted to my supervisor Dr C.S. Nelson of the Earth Science Department, University of Waikato, for his encouragement and help that was freely given during all phases of this study and for his constructive criticism of the manuscript. Dr Nelson also supplied several of the photographs. Thanks are also expressed to members of staff, particularly Professor J.D. McCraw, and colleagues at the University who assisted in this study, and to the University of Waikato for financial support in the form of a University of Waikato Post-Graduate Scholarship. The New Zealand Oceanographic Institute kindly supplied samples of western shelf sediments; the Meat Research Institute, Hamilton, provided the transmission electron microscope facilities and the Engineering Department, University of Auckland and University of Waikato provided the scanning electron microscope facilities. The author wishes to thank Mrs M. Henrickson and Heather Kunze for their excellent typing and Mr J. Julian for printing the photographs. I would like to record sincere thanks to my wife Anne for her assistance in the field and in the compilation of this manuscript, and for her support, both financial and moral, that helped make this study possible.

CHAPTER 1

INTRODUCTION

1.1 THE SIGNIFICANCE OF CLAY PETROLOGY IN THE EARTH SCIENCES.

Clay minerals are major constituents of mudstones and related argillaceous rocks, which together constitute some 75% of sedimentary deposits, and occur in significant amounts in a further 20% of sediments (cf. Weaver, 1958a, p.254; Picard, 1971). This volumetric abundance implies that the study of clay minerals and clay-rich sediments is of vital importance in understanding earth processes. The development of clay mineralogy as a science has grown from the need to understand the physical and chemical factors controlling the properties of this economically important group of sediments. Such properties determine the agricultural potential of soils and the ceramic qualities of clay materials, they have a significant influence on the engineering properties of sediments and on the cap rock and reservoir properties in petroleum geology, and they provide information on the nature and position of hydrothermal ore deposits. Clay geology evolved from these essentially economic roots and led to detailed investigations of areal and stratigraphic variations in clay mineral assemblages with several major aims:

(1) to establish relationships between clay mineral species and environments of deposition;

(2) to attempt to establish criteria for distinguishing between detrital and authigenic clay minerals;

(3) to assist in paleogeographic and provenance studies;

(4) to provide a tool for assisting stratigraphic correlation;

(5) to investigate the location and economic potential of clay minerals and clay deposits.

Many workers have shown that clay mineral distributions vary systematically with physical and chemical differences in sedimentary environments, both ancient and recent, although in many instances the interpretation of these distributions is equivocal. However, Weaver (1958b) has demonstrated clearly that when clay data are combined with petrographic, paleontologic and field evidence, significant geologic conclusions can be made.

1.2 AREA OF STUDY AND GEOLOGIC BACKGROUND.

This account presents the results of a detailed study of the clay mineralogy of Mesozoic to Recent sediments in the South Auckland to Wanganui region and of Recent sediments from the adjacent western continental shelf of the North Island, New Zealand, a combined total area of about 70,000 km^2 (Fig.1.1). The onshore study area includes most of the Waikato, King Country, Taranaki and Wanganui Land Districts. For convenience, this area is divided into the Hamilton, Taranaki and Wanganui Districts (Fig.1.1), these being equivalent to the areas covered by the 1:250,000 New Zealand Geologic Survey maps of Lensen et al. (1959), Kear (1960) and Hay (1967), respectively. Place names and major physiographic features referred to in this thesis are presented in Fig.1.2. The distribution of the major sedimentary units in the onshore study area is described in

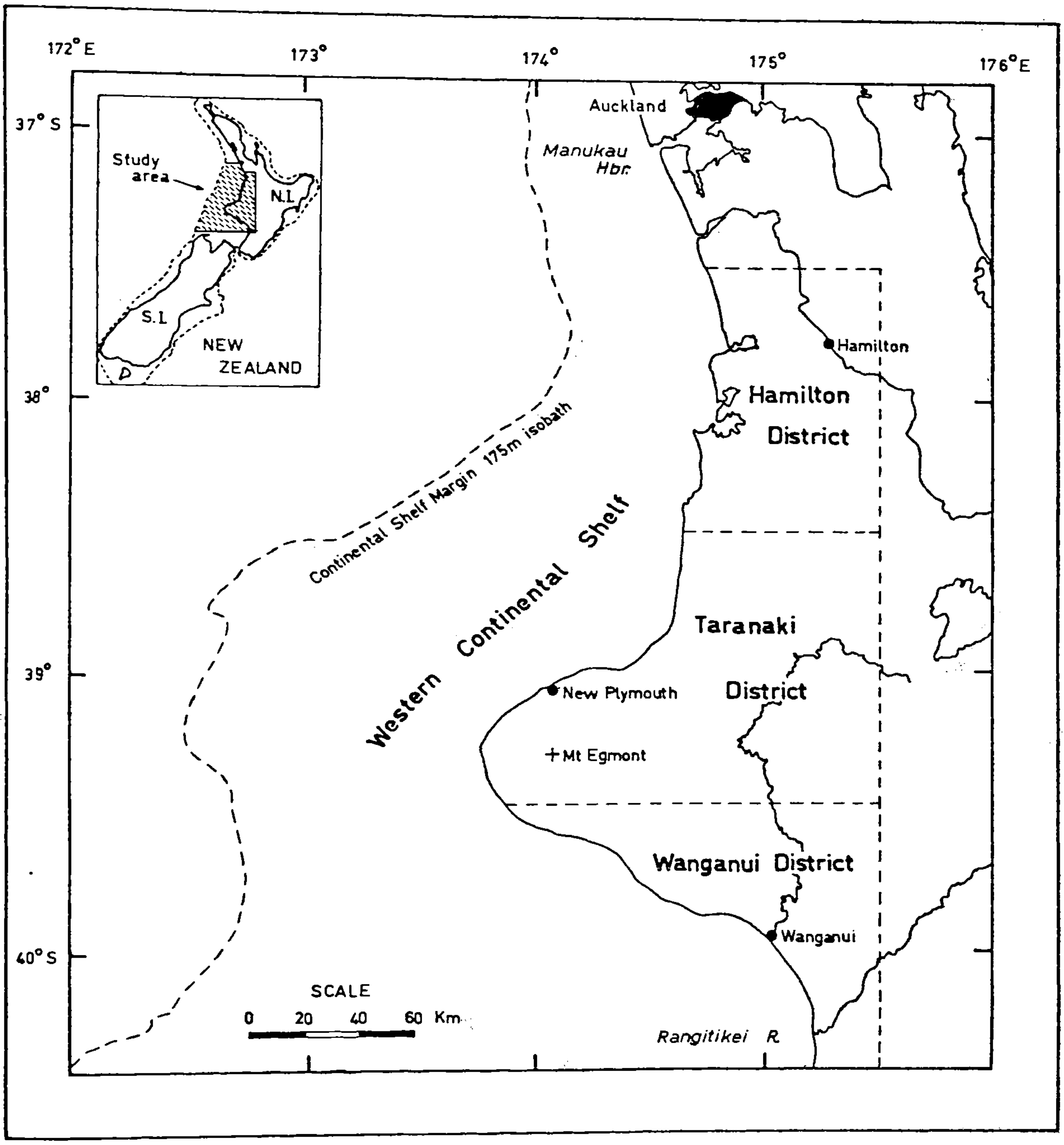


Fig.1.1. Area of Study

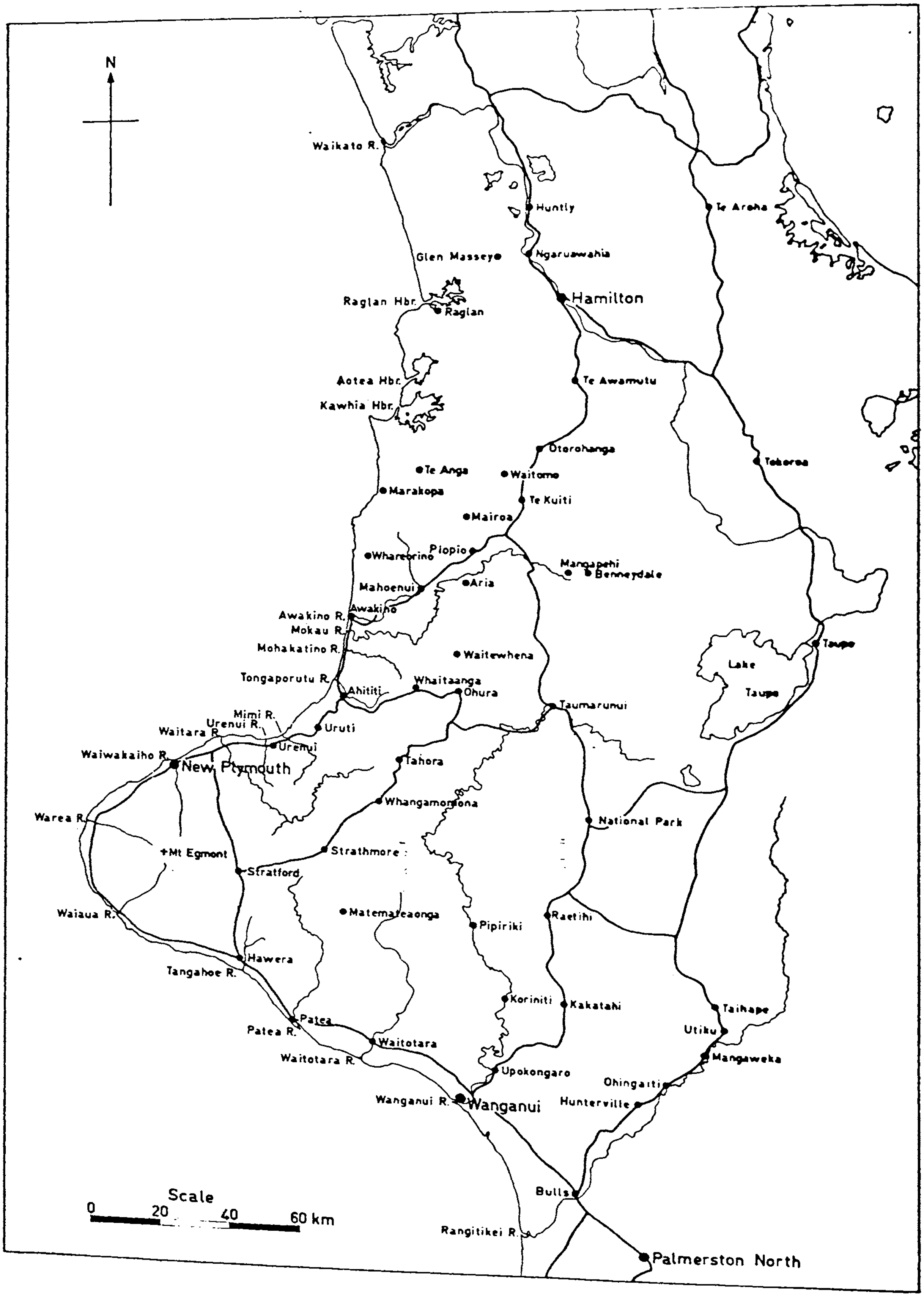


Fig. 1.2. Major place names, physiographic features and roads in the study area. Additional "lift-out" copy available in back pocket.

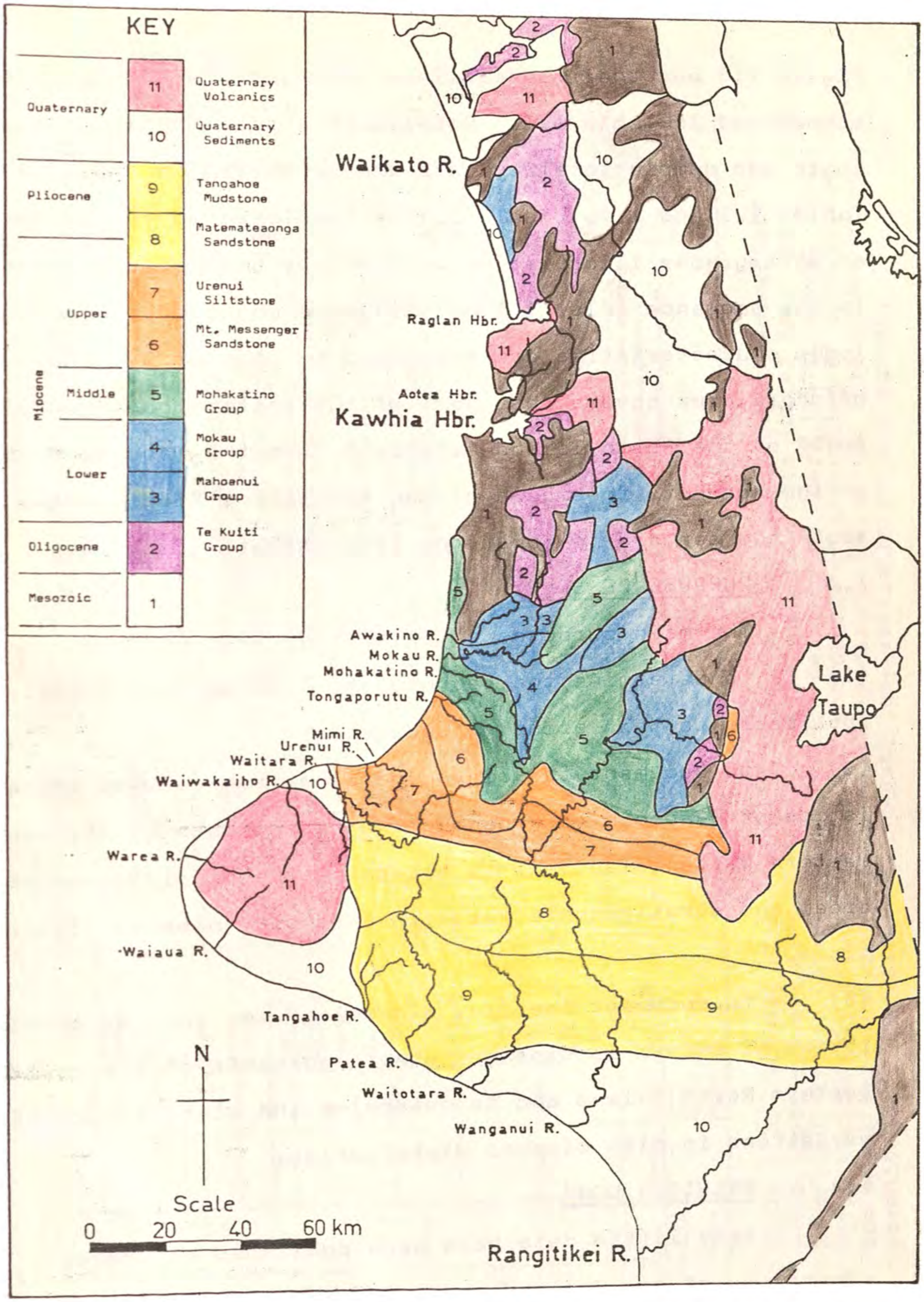


Fig. 1.3. Simplified geologic map of the onshore study area (after Lensen et al., 1959; Kear, 1960; Hay, 1967). Additional "lift-out" copy available in back pocket.

Figure 1.3 and their generalised stratigraphy and lithology summarised in Table 1.1. Details of stratigraphic, lithologic and sample locality data are presented in Appendix Tables 1.1 and 1.2. Table 1.1 is complemented by a glossary of photographs illustrating most of the sedimentary units in the sequence (Pls. 1.1 - 1.24) more detailed sedimentologic characteristics are discussed in Chapter 3. The offshore area covers that part of the western continental shelf of the North Island extending from the Manukau Harbour in the north to the mouth of the Rangitikei River in the south and seaward to about the 175m isobath (Fig.1.1).

1.3 PURPOSE OF STUDY

The main purposes of this study are:

(1) to determine the techniques of study most readily applicable to this group of sediments;

(2) to document the clay mineralogy and related characteristics of Mesozoic to Recent onshore sediments of the central Western North Island and to determine the significance of areal and stratigraphic variations in clay mineral distributions.

(3) to document the clay mineralogy and related characteristics of modern continental shelf sediments of the central Western North Island and to determine the significance of variations in clay mineral distribution.

1.4 PREVIOUS WORK

Very little data have been published on the clay mineralogy of New Zealand rocks and sediments. Most publications have concerned themselves with the mineralogy of specific economic clay deposits, or have dealt with clay

STRATIGRAPHY						GENERALISED LITHOLOGY	Plate Reference
Period	Epoch	Series	N.Z.Stage	Group	Formation		
Quaternary	Recent (Rc)					Volcanic ash and soils.	-
	Pleistocene	Hawera (H)				Silt, sand, gravel and volcanogenic debris.	-
		Wanganui	Castlecliffian (Wc)			Siltstone, sand, conglomerate and shell beds.	-
			Nukumaruan (Wn)			Limestone, siltstone, sandstone, shell beds and lignite bands.	-
Tertiary	Pliocene		Wo - Wm		Tangahoe Mudstone (Tg)	Massive blue-grey mudstone with some sandstone and concretions.	1.23, 1.24
			Tk - Wo		Matemateaonga Sandstone (Mt)	Blue-grey mudstone, sandstone, conglomerate, shell beds and some concretions.	1.22
	Up.Miocene	Taranaki	Up. Tt		Urenui Siltstone (Ur)	Micaceous blue-grey siltstone with some concretions.	1.21
			Lr.-Mid. Tt		Mt.Messenger Sandstone (Mt.M)	Massive and bedded sandstone and lesser mudstone, slightly tuffaceous, with some coal beds and slump structures.	1.19, 1.20
			Lr. Tt	Mohakatino	Tawariki Mudstone (Tw)	Blue-grey mudstone and sandstone tuffaceous in part.	1.18
	Mid.Miocene	Southland	Sw		Ferry Sandstone (Fy)	Non-tuffaceous sandstone with some mudstone and slump structures.	1.16, 1.17
			Sw - Lr.Tt		Omoso Formation (Om)	Alternating beds of micaceous sandstone and siltstone with minor volcanic detritus.	-
			Sc - Sw		Purupuru volcanic Sandstone (Pu)	Tuffaceous sandstone and mudstone interbedded in part.	1.14, 1.15
	Lr.Miocene	Pareora	Mid.-Up. Pl		Mangarara Sandstone (Mg)	Tuffaceous calcareous sandstone with some limestone.	1.16 1.14
			Mid. Pl	Mokau	Upper Mokau Sandstone (U.Mo)	Blue-grey argillaceous fine sandstone with some conglomerates and concretions.	1.14
			Mid. Pl		Maryville Coal Measures (My)	Upper and lower coal zones and carbonaceous shale separated by carbonaceous sandstone.	-
			Lr.-Mid.Pl		Lower Mokau Sandstone (L.Mo)	Massive, ferruginous, argillaceous sandstone, carbonaceous or thin coal seams.	1.11, 1.12 1.13
			Po - Pl	Mahoenui	Taumarunui Formation (Tmr)	Flysh-type graded beds of alternating sandstone and mudstone.	1.10
			Up.Lw - Po		Taumatamaire Formation (Tmt)	Massive blue-grey mudstone with some sandstone and limestone.	1.9
	Oligocene	Landon	Lw	Te Kuiti	Otorohanga Limestone (Ot)	Crystalline flaggy limestone.	1.7, 1.8
			Ld - Lw		Waitomo Sandstone (Wt)	Massive calcareous sandstone.	1.7
			Ld - Lw		Orahiri Limestone (Or)	Sandy limestone, flaggy in part.	1.6, 1.7
			Ld		Mangapehi Sandstone (Mp)	Calcareous, glauconitic sandstone.	-
			Mid.Lwh-Ld		Te Akatea Siltstone (T.A)	Calcareous mudstone.	-
			Mid.Lwh-Ld		Aotea Sandstone (Ao)	Calcareous sandstone with some mudstone and limestone.	1.3, 1.5
			Lr.-Mid.Lwh		Whaingaroa Siltstone (Wh)	Massive blue-grey, calcareous siltstone with occasional limestone.	1.7
			Lwh		Glen Massey Sandstone (G.M)	Calcareous mudstone, sandstone and limestone.	1.3, 1.4 -
	Up.Eocene	Arnold	Ar - Lwh		Mangakotuku Siltstone (Mk)	Carbonaceous, purple-brown, non-calcareous siltstone with some concretions.	-
			Ar - Lr.Lwh		Waikato Coal Measures (Wk)	Carbonaceous mudstone, coal beds and some sandstone and conglomerate.	1.2
Jurassic		Otoke Kawhia, Herangi				Slightly indurated mudstone, sandstone and conglomerate with some tuff.	-
Triassic		Balfour				Indurated mudstone and sandstone with some tuff.	1.1

Table 1.1. Summary of the stratigraphy and lithology of major sedimentary units (note abbreviations in brackets) in the study area (after Kear and Schofield, 1959; Lensen <u>et al</u>., 1959; Kear, 1960; Hay, 1967; Happy, 1971). Where sedimentary structures, colour and degree of induration of lithologies are specified, the property is generally diagnostic of the bulk of the strati-graphic unit. Additional "lift-out" copy available in back pocket.

pedology. The clay mineralogy documented for rocks, tephras and soils of the region is summarised below.

It is relevant to summarise briefly the petrographic character of Mesozoic basement rocks which represent the primary source of terrigenous material for Cenozoic sediments in the study area. The range in composition of Mesozoic rocks is large. Martin (1967) showed that Mesozoic basement sandstones of the Oparau Facies in the western Hamilton District (Figs. 1.1 and 1.3) are rich in acid andesine plagioclose and volcanic (mainly andesitic) rock fragments set in a conspicuous argillaceous matrix. In contrast, the Mesozoic basement sandstones of the Morrinsville Facies in the eastern Hamilton District (Figs. 1.1 and 1.3) are dominated by volcanic and other rock fragments and quartz, with lesser oligoclase-andesine plagioclase and little matrix (Finlow-Bates, 1970). Nelson (1973) showed that the bulk mineralogy of Mesozoic sandstones and mudstones in the Waitomo County is dominated by clay minerals (av. 30-45%), plagioclase (av. 25-35%) and quartz (av. 20-25%), although the mudstones may contain considerably greater quantities of clay minerals than the suggested average. In general the Mesozoic sandstones and mudstones have a clay fraction dominated by illite and chlorite and by mixed-layer clays, with lesser montmorillonite, muscovite and biotite (Fieldes et al., 1968; Nelson, 1973).

Nelson (1973), working in the South Auckland area, presented a comprehensive account of the clay minerals in the Oligocene Te Kuiti Group sediments and some clay data on the underlying Mesozoic and overlying Mahoenui Group rocks (Fig. 1.4). He considered that the small to moderate

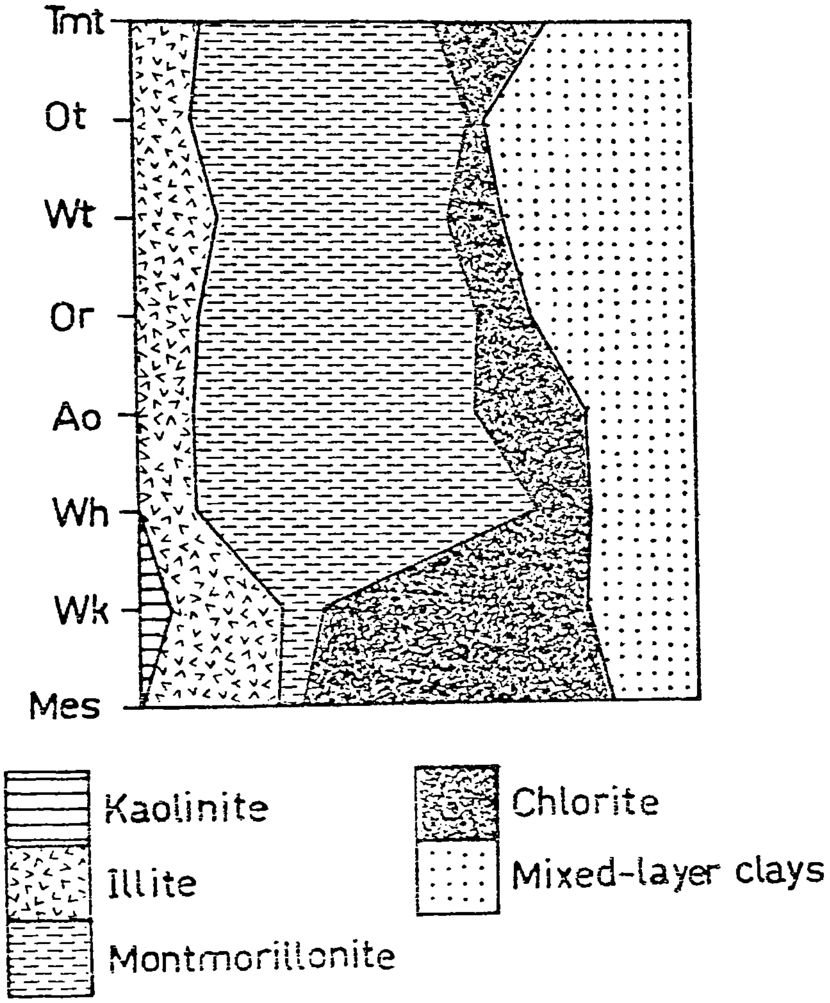


Fig. 1.4. Relative clay mineral abundance in the $<4\mu$ size fraction of Mesozoic, Te Kuiti and Mahoenui Group sediments of the Waitomo County (after Nelson, 1973). Mes = Mesozoic sandstones and mudstones; Wk = Waikato Coal Measures; Wh = Whaingaroa Siltstone; Ao = Aotea Sandstone; Or = Orahiri Limestone; Wt = Waitomo Sandstone; Ot = Otorohanga Limestone; Tmt = Taumatamaire Formation mudstones.

quantities of illite and chlorite in the Te Kuiti Group are probably mainly detrital, being derived directly from the Mesozoic sandstones and mudstones. The small amounts of kaolinite in the Waikato Coal Measures sediments were regarded as a product of acid leaching associated with carbonaceous material. The scarcity of montmorillonite in the Mesozoic rocks precluded a direct detrital origin for the abundant montmorillonite in the Te Kuiti Group but Nelson considered that (p.440) "... substantial amounts may have been inherited from Oligocene soils ...", or (p.441)"... may have originated by the aggradational transformation of several of the soil clays", and that some was possibly neoformed. Where reference to Nelson's work is made in this thesis it is important to keep in mind that his clay mineral analyses were conducted on the <4μ size fraction whereas the <2μ fraction was used most commonly in this work.

Ker (1973) described the stratigraphy and engineering geology of the lower Wanganui Valley as part of a survey of a proposed hydro-electric power development on the Wanganui River. He found that the Upper Pliocene to Lower Pleistocene sandstones and siltstones in this area are characterised by the presence of abundant chlorite and sericite and the virtual absence of expandable clay minerals.

About 70% of the study area is mantled with Quaternary rhyolitic and andesitic tephras derived from eruptive centres in the central North Island and Taranaki areas. Loess is a common component of these beds and paleosol formation is widespread (Gibbs, H.S., 1968; Symes and Wells, 1973; Kirkman, 1975). The tephras derived from the central North Island eruptives have a clay mineralogy dominated by allophane in Holocene units (i.e. <10,000 years in age) and by halloysite in Late Pleistocene units (i.e. <42,000 years in age). The observed distribution of allophane and halloysite in the Holocene and Late Pleistocene tephras suggests the following weathering sequence operates: feldspar and glass ⟶ allophane ⟶

poorly crystallised halloysite ⟶ well crystallised halloysite ⟶ metahalloysite. (Kirkman, 1975).

By far the most commonly developed soil materials on Mesozoic and Tertiary sediments are the yellow-brown earths (Fieldes et al., 1968). This group of soils forms under a wide range of climatic and leaching conditions and the clay fraction of the different soils is complex, comprising variable amounts of chlorite, illite, vermiculite, montmorillonite, kaolinite and mixed-layer clays. While the composition of the soils is dependent initially on the composition of the parent material, with increasing weathering intensity and time soils with a high kaolinite content are typical (Fieldes et al., 1968). Mesozoic sandstones and mudstones have sometimes been subjected to varying degrees of weathering for a considerable period of time, and the often very deep red weathered profile, up to 30m thick, is indicative of an advanced state of alteration. These thick sections of weathered rocks contain a variety of clay minerals intermediate between those in the parent rocks and those in the soil, but kaolinite is generally dominant (Fieldes et al., 1968). In the South Auckland region, clay minerals in surface soils, as well as in buried paleosols derived from central North Island rhyolitic and andesitic tephras, are dominantly allophane, (in tephras < 10,000 years in age) and halloysite and/or kaolinite (in tephras >10,000 years in age). Lesser amounts of montmorillonite and micaceous clay minerals are also present (Fieldes, 1968; Pullar et al., 1973; Kirkman, 1976). The similarity of clay mineralogy between paleosols and parent tephras indicates that during the comparatively short period (generally < 2,000 years) between tephra deposition and subsequent burial by later ash or loess the paleosols have been only weakly weathered. Loams and sandy loams developed on the younger (< 12,000 years in age) andesitic Mount Egmont derived ashes in the southwestern Taranaki District are composed of subequal amounts of allophane and hydrous feldspar (Fieldes, 1968; Symes and Wells, 1973).

1.5 TERMINOLOGY

Several terms used in this thesis are applied with various meanings in the literature. Accordingly, the definitions used here are defined below so as to avoid any ambiguity.

Clay Size. Clay-sized particles are those particles having < 2μ equivalent spherical diameter (see Carrol, 1970, p.2) for a discussion of the upper size boundary; equivalent spherical diameter, or e.s.d, is a measure of particle size, equal to the computed diameter of a hypothetical sphere of specific gravity 2.65 (i.e. quartz) having the same settling velocity and same density as those calculated for a given sedimentary particle in the same fluid).

Clay Mineral. Clay minerals are crystalline, hydrous silicates with layer or chain lattices consisting of Si-O tetrahedral sheets arranged in hexagonal form and condensed with octahedral layer (Al-O-OH). Each sheet has planes of cations (Si, Al, Mg, Fe) with each surrounded by either four (tetrahedral) or six (octahedral) O and OH ions. Mixed-layer clay minerals are either randomly or regularly interstratified intergrowths of two or more clay minerals. (Carroll, 1970).

Detrital Inheritance. The process whereby clay minerals are derived from a landmass and remain unmodified because of their stability. Such clay minerals are detritally inherited, or simply detrital clays (Millot, 1970).

Transformation. The process whereby clay minerals undergo lattice changes by processes acting within the environment which may involve the subtraction (degradation) or addition (aggradation) of ion constituents. Such clays are transformed clays (Millot, 1970).

Neoformation. The process whereby clay minerals are in large part or wholly precipitated from products dissolved in the hydrosphere. In leaching environments clay mineral neoformations result from the subtraction of ions during the sericitization of feldspars, chloritization of ferromagnesian minerals, evolution of volcanic glass into montmorillonite and the kaolinization of crystalline rocks. In confined environments clay mineral neoformations occur through the recombination and addition of ions to form products such as montmorillonite, attapulgite and sepiolite. Such clays are neoformed clays (Millot, 1970).

Authigenesis. The creation of minerals *in situ* by chemical reactions in sediments at the time of, or following deposition (Weller *et al*., 1960). Both transformed and neoformed clays can be authigenic.

CHAPTER 2

TECHNIQUES OF STUDY

2.1 INTRODUCTION

Clay mineral investigations, especially quantitative analysis, are hampered by the fine size and diverse chemistry and structure of clays. Consequently a large volume of literature exists concerning techniques of clay mineral analysis, especially those utilising X-ray diffraction (XRD), but inevitably the recommendations are not entirely satisfactory for all sediment types (cf. Thorez, 1975, pp. XV-XVI). In this study the clay mineralogy of a large number of samples of widely differing ages and of diverse texture and composition has been investigated. Accordingly, preliminary experimentation was necessary to establish methods of analysis that were not only rapid and reproducible, but also applicable to most samples. This chapter details the various techniques attempted and finally used in this study, in the anticipation that the information will be useful to other workers embarking on similar studies.

2.2 SAMPLING PROCEDURES

2.2.1 ONSHORE SAMPLES

One hundred and seventy-two samples of Mesozoic to Recent sediments were collected using a hammer and chisel or spade from 88 localities (Fig. 2.1) in the Hamilton, Taranaki and Wanganui Districts. Considerable care was taken to avoid sampling obviously weathered material. Stratigraphic, lithologic and sample locality data are presented in Appendix

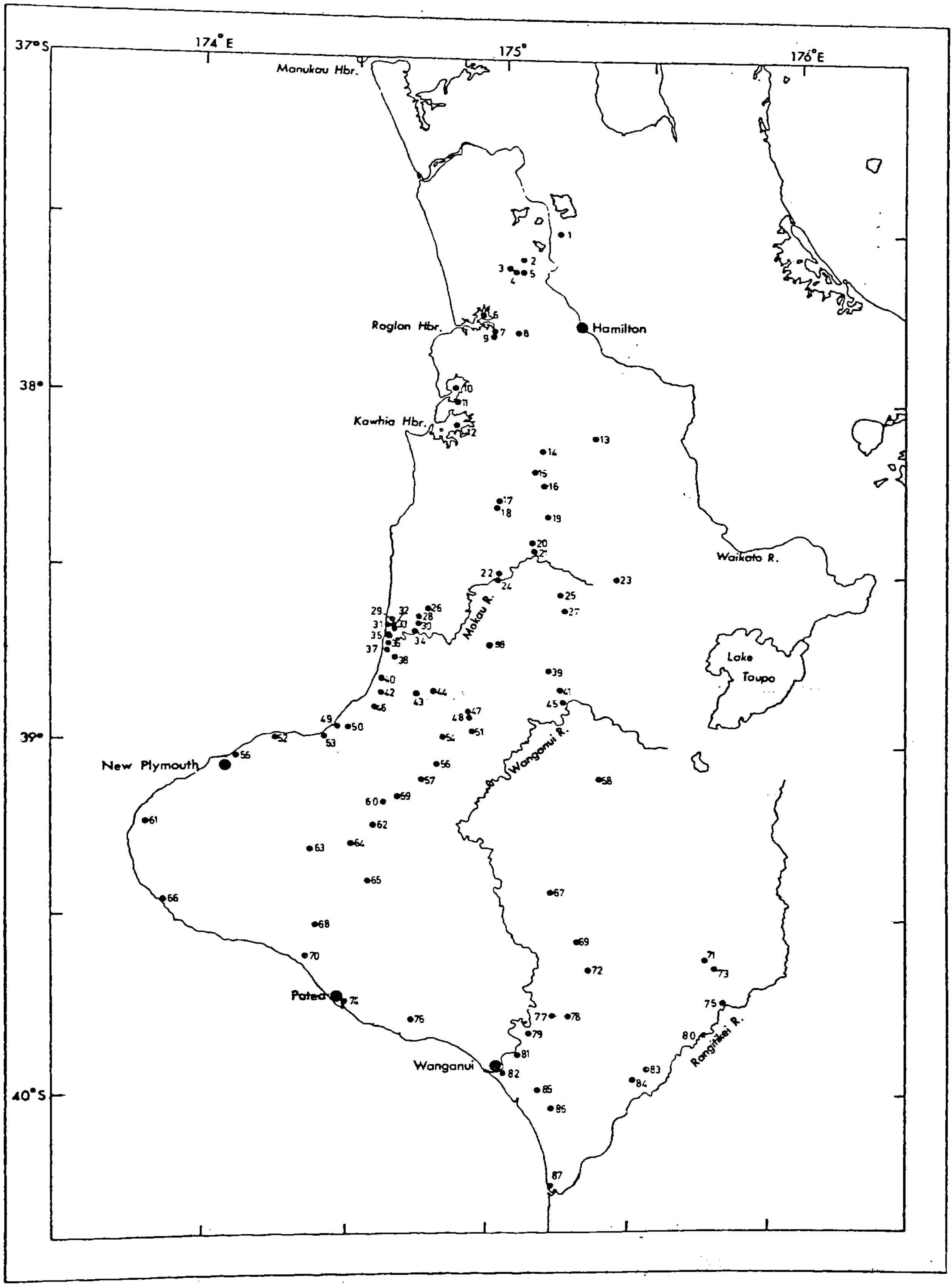


Fig. 2.1. Sample localities of onshore and harbour and river samples. Additional "lift-out" copy available in back pocket.

Tables I.1 and I.2 and correlation of thesis sample localities and sample numbers with those used during field work is given in Appendix Table II.1. Each sedimentary formation was sampled at several locations, including at or near type and/or reference sections, and samples were taken from the various lithologic sub-units in the formation. The density of sampling was governed by three factors: (1) The number of formations in the study area; (2) the necessity to sample formations in sufficient density to detect any major vertical or lateral variations in clay mineralogy between samples from the same formation; and (3) the time available for study. The results of all analyses were tabulated in stratigraphic order (Appendices III-IX). This permitted ready comparison of variations in mineral abundances in samples between and within formations and between different localities. Initial results showed that within formations the qualitative clay mineralogy of samples was generally similar, but that there could be marked differences in quantitative clay mineralogy that were largely a function of lithology. Reasons for these differences are discussed elsewhere (pp. 91-94). It became apparent that the clay mineralogy of mudstones in formations represented best the average formational mineralogy. Within the limits of quantitative clay mineral interpretation (i.e. ±10%, see pp. 51-55) there exists only small variations in clay mineral abundance between mudstone samples from the same formation. On this basis the sampling density achieved was considered adequate, and trends in clay mineral abundance plotted from formational average data are considered to represent realistic differences or similarities in clay mineral suites between formations. The writer realises that more detailed sampling might produce different values for formational averages, but predicts that these differences would be small and alter only slightly the major trends in the clay mineral distribution patterns obtained.

2.2.2 OFFSHORE SAMPLES

Sixty-eight surficial sediment samples from the western continental shelf (Fig. 2.2, A-C prefix numbers) were provided by the New Zealand Oceanographic Institute. The location and depth of samples, the recovery methods, and the texture of the sediments are described by McDougall and Brodie (1967, Table 1). Of the 48 samples used in clay mineral analysis in this study, 3 were collected with an Agassiz trawl, 6 with a cone dredge and 39 with either a Petersen or a Dietz grab sampler. The New Zealand Oceanographic Institute also provided 6 surface seston (suspended solids) samples from western continental shelf locations off Raglan Harbour and Cape Egmont (Appendix Table I.3; Fig.2.2, N prefix numbers). An additional 24 surficial sediment samples were collected by the writer in August 1973 from tidal flats of harbours and river mouths bordering the shelf using a plastic scoop (Appendix Table I.4, Figs. 2.1 and 2.2). The samples from Raglan Harbour were collected by Sherwood (1973) using a 500cc capacity Marukawa grab. With the exception of the Raglan Harbour samples, the harbour and river samples were analysed within a few weeks of their recovery and while still damp.

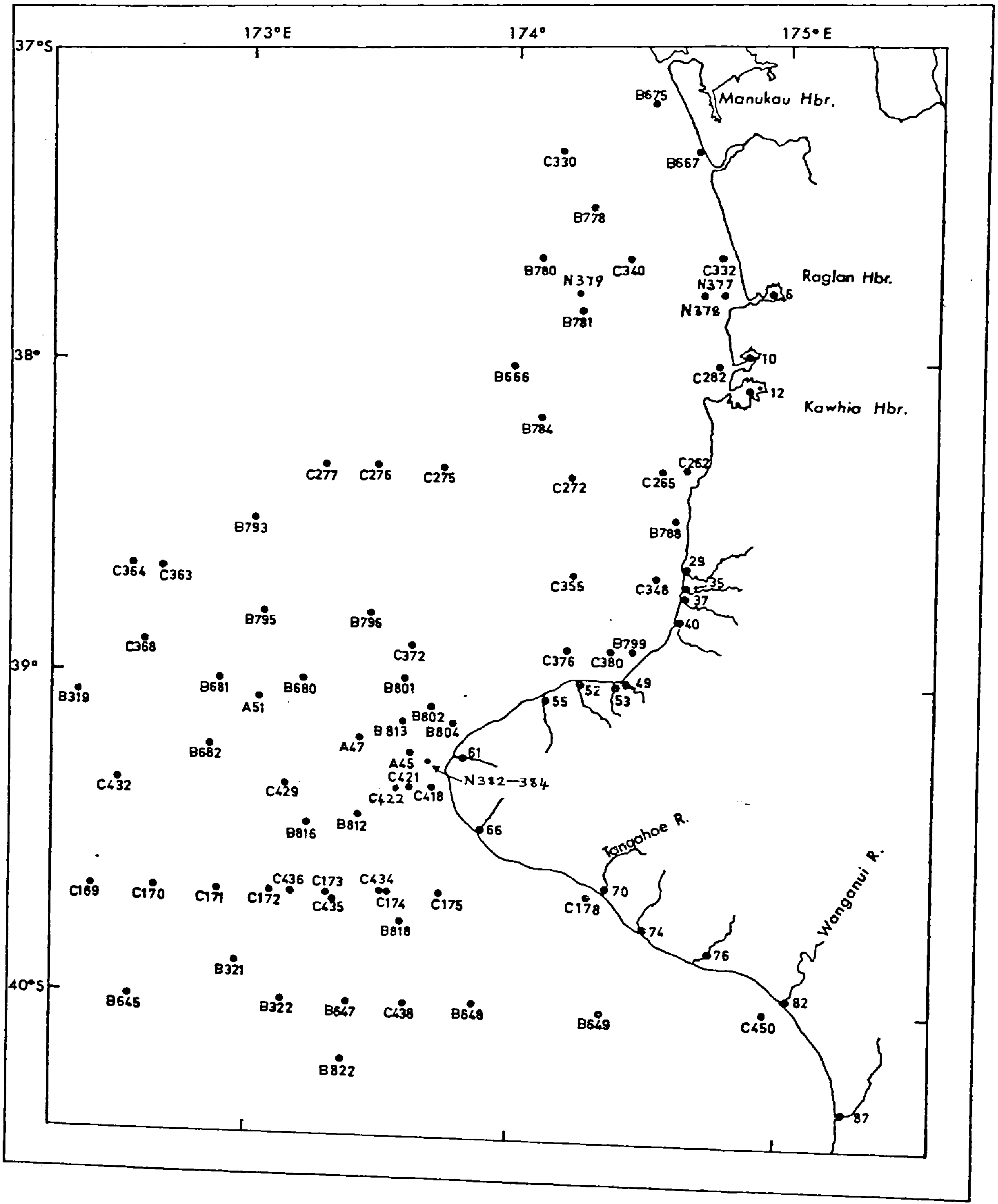


Fig. 2.2. Sample localities of western shelf and harbour and river samples. Additional "lift-out" copy available in back pocket.

The use of various types of sediment samplers and the analysis of samples in different states of dryness impose some restrictions on the comparative interpretation of clay mineralogic data (Siegel and Pierce, 1973). Some workers have reported no appreciable effects on the clay mineralogy of the $< 2\mu$ size fraction as a result of drying of samples (Griffin, 1962, p.740; Griffin and Goldberg, 1963, p.729). However, Biscaye (1965, p.807) did observe "...that a partly expandable, vermicultite-type mineral present in the wet samples of the $2 - 20\mu$ size fraction may have altered to chlorite upon drying of the core material". By comparing clay mineral contents of Recent marine sediments collected as piston and gravity cores and grab samples, Siegel and Pierce (1973) have shown that care must be taken in making comparisons of semiquantified clay mineral abundances in Recent marine sediments if the same sampling device has not been used to extract bottom sediment. Such differences in the clay mineral suites may result from selective washing out of clay minerals during sampler recovery, or may be due to differences in sediment sampling action. The majority of offshore sediments analysed in this study were collected with grab samplers. One would expect those samples collected with trawl and dredge equipment to be less representative of "true" bottom sediment mineralogy because of possible washing out of "fines" during sampler recovery. On the other hand the river and harbour samples, collected with a plastic scoop, would be most representative of "true" bottom sediment composition. Time and economic factors prevented collection of samples with standard gear.

2.3 ANALYTICAL PROCEDURES

2.3.1 FRACTIONATION PROCEDURE

Steps involved in the preparation of samples for both bulk compositional and clay mineral analysis are summarised in a flow chart (Table 2.1). Details of sample preparation for specific instrumental analysis are discussed in the following section.

2.3.2 EXAMINATION PROCEDURES AND INSTRUMENTAL TECHNIQUES.

2.3.2A X-RAY DIFFRACTION

X-ray diffraction proved to be the most useful and the most rapid method of mineral analysis. XRD techniques were used both for determining the modal mineral composition of sediments and for clay mineral analysis.

BULK AND MODAL MINERAL ANALYSIS

The XRD modal analysis method of Nelson and Cochrane (1970) was used for the quantitative determinations of quartz, plagioclase, potash feldspar and clay minerals (Appendix Tables III.1 and III.2.). Because $CaCO_3$ occurs as both aragonite and calcite in some samples, $CaCO_3$ was determined by chemical methods. Apart from the initial standardisation of procedure and equipment the XRD modal analysis technique is very rapid (about 7 minutes per sample), and although the estimated accuracy is only about ± 10% it is considered to provide reliable indications of general mineral trends in samples (Nelson and Cochrane, 1970). Many of the sediments studied were very fine-grained and conventional point-counting techniques would have been both inadequate and subject to large errors. Moreover, many

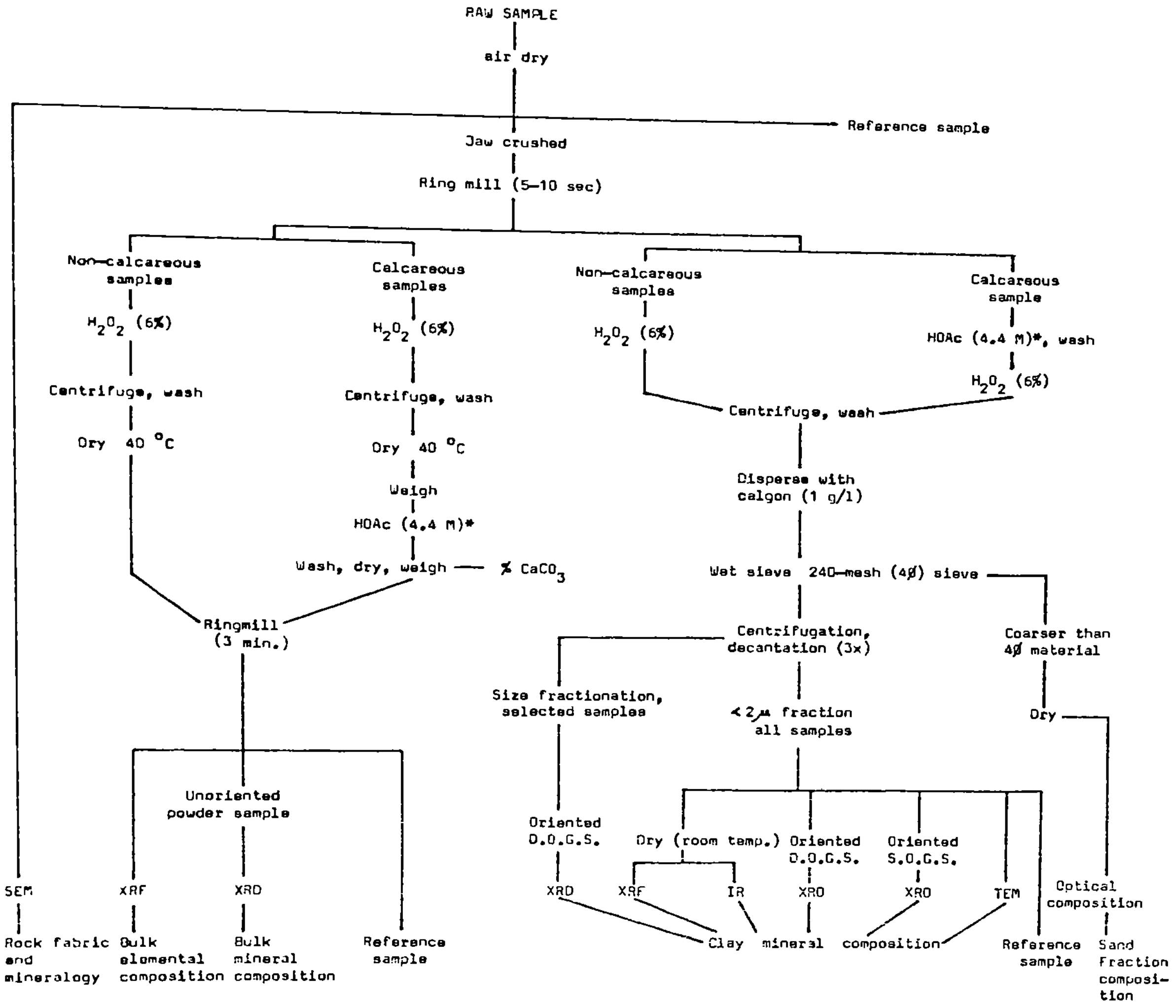


Table 2.1. Flow chart summarising steps in the preparation of samples for bulk compositional and clay mineral analysis. Clay minerals were mounted for X-ray diffraction analysis using the dropper-on-glass slide (DOGS) and the smear-on-glass slide (SOGS) techniques. Analyses were carried out employing the following methods: X-ray diffraction (XRD), X-ray fluorescence (XRF), infrared absorption (IR), transmission electron microscopy (TEM), and scanning electron microscopy (SEM).

+ Samples were milled for 5-10 seconds to facilitate acid digestion. Preliminary tests showed that crushing for 5-10 seconds in the ring mill did not alter the distribution of clay mineral species in the various size fractions. After milling, samples were split by pouring into a cone and quartering.

* Carroll and Starkey (1971) have shown that clay minerals are not altered by treatment with acetic acid if solution strengths are less than 4.4M.

samples were unconsolidated or friable and contained swelling clays, which would have made thin-section preparation both difficult and time-consuming.

Sample Preparation and Analysis

The quantitative procedures were carried out on a Philips P.W. 1050 geiger-counter X-ray diffraction spectrometer using nickel-filtered copper radiation at the following machine settings:

KV	36
MA	16
Time constant	4
Attenuation	2
Chart speed	20 mm/min.
Scanning speed	2° 2θ/min.
High voltage	56%
Window level	L —— ∞
Lower level	17%
Oscillator	high
Slits - divergence	1°
- scatter	1 mm
Ratemeter setting	4 or 10

Initial operating conditions were standardised using a method similar to that suggested by Nelson and Cochrane (1970, p.152). In this study, however, a permanent quartz standard was prepared by grinding a pure sample of quartz to less than 350-mesh (45µ) particle size, mixing the powder with araldite resin in a Philips powder sample holder and when set, grinding the mixture flush with the sample holder. The standard mineral powder mixtures used by Nelson and Cochrane (1970) were obtained from the Department of Geology, University of Auckland. The mixtures were run from 18° 2θ to 32° 2θ at a ratemeter setting of 400 cps (RMS 4); an additional ratemeter setting of 1000 cps (RMS 10) was employed when peaks under analysis were off-scale.

Interpretation of Data

A straight regional background line is drawn on each diffractometer trace between 18° and 32° 2θ. The vertical distance between peak apex and background line, expressed as the number of small grid units on the diffractometer trace to the nearest half unit, is taken as equal to the peak height intensity of the mineral under analysis. Reflections used for measuring peak height intensities above background are given in Table 2.2.

Mineral	Diffractometer Setting	Peak Position	
		°2θ	dÅ
Quartz	RMS 4	20.8	4.26
Plagioclase	RMS 4	28.0	3.20
Potash feldspar	RMS 4	27.5	3.25
Calcite	RMS 4	29.4	3.04
Clay minerals	RMS 4	19.9	4.46
Quartz	RMS 10	26.6	3.34
Plagioclase	RMS 10	28.0	3.20

Table 2.2. Reflections used for measuring XRD peak height above background at selected instrument settings.

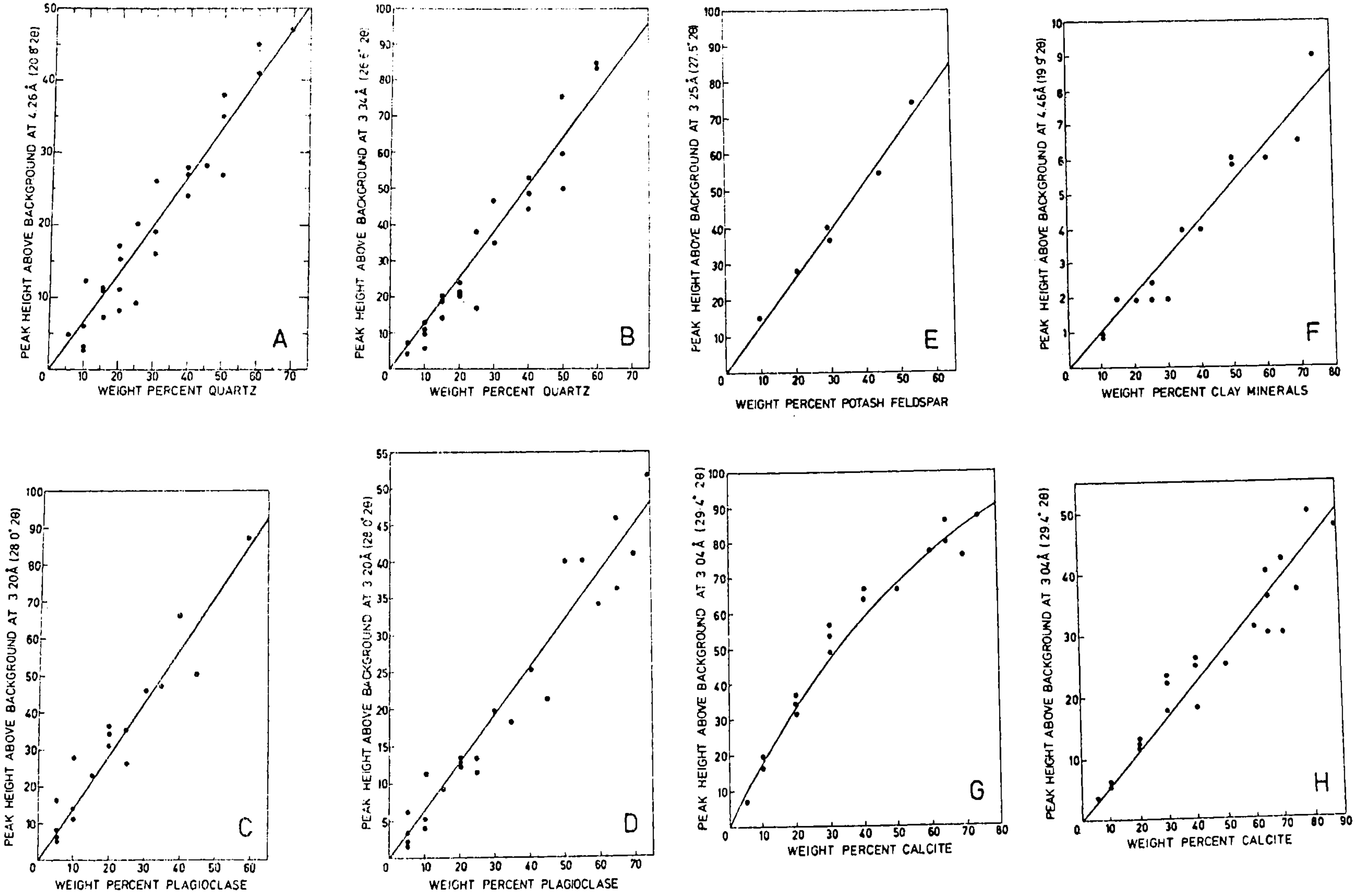


Fig. 2.3. Intensity concentration curves for quartz (A at RMS 4; B at RMS 10), plagioclase feldspar (C at RMS 4; B at RMS 10), potash feldspar (E at RMS 4), clay minerals (F at RMS 4) and calcite (G at RMS 4; H at RMS 10). The linear correlation coefficient (at the 95% significance level) for each of the intensity-concentration plots of each mineral are as follows: A = 0.989; B = 0.985; C = 0.985; D = 0.989; E = 0.998; F = 0.991; and H = 0.986.

Carlson and Rogers (1973) suggested that more accurate results would be obtained by measuring peak area rather than peak height. However in this study peak area measurements were made and they did not improve the accuracy of the method so that the simpler peak height measurements were adopted. The peak height intensities above background were plotted against the weight percent of minerals in the 44 standard powder mixtures. The data suggest that an approximately linear relationship exists between peak height and concentration for quartz, plagioclase, potash feldspar, clay and calcite (at RMS 10), but a parabolic curve best fits the data for calcite at RMS 4 (Fig.2.3 G). Statistical analysis shows that plots A to F, and H, are best represented by straight lines and that the data correlate well with these calculated lines (Fig.2.3). A log/log plot of calcite at RMS 4 shows that a simple parabolic curve does not fit the data well; hence a freehand line was drawn through the points. Nelson and Cochrane (1970, p.161) suggested that the accuracy of the technique is about ± 10% of the true modal value but did not mention the basis on which this figure was derived. For the standard powder mixtures the accuracy of the technique obviously varies with mineral type and machine setting (Fig.2.3), and it is reasonable to assume that this will be the case for rock 'unknown' determinations. The standard errors in mineral concentration determined statistically for each of the linear intensity-concentration plots (Fig.2.3) are: quartz (RMS 4) = ± 4.2%; quartz (RMS 10) = ± 4.4%; plagioclase (RMS 4) = ± 3.0%; plagioclase (RMS 10) = ± 4.6%; potash feldspar (RMS 4) =

± 3.4% and calcite (RMS 10) = ± 6.6%. These data show that the standard errors are less than ± 5% except for calcite, and give an estimate of error for the unknown mineral quantifications. The standard error in mineral concentration represents the average deviation of the points from the best fit intensity-concentration line in the x-axis direction and is not a maximum deviation. However, for reasons discussed in the remainder of this section, the errors in certain individual mineral concentration determinations are probably larger than those suggested by the standard errors and makes calculation of errors for individual minerals difficult. Because of this Nelson and Cochranes' (1970) estimate of error of ± 10% for the technique is considered to be of the right order of magnitude and is probably nearer the upper limit of error for the technique.

Problems imposed by the quantity of iron in samples (Nelson and Cochrane, 1970, p.153) were overcome by using a lower level window filter set at 17%. This reduced iron fluorescence to such a level that it caused no appreciable differences in background height on diffractometer traces.

Nelson and Cochrane (1970) outlined some of the problems inherent in this technique, two of which are important in this study. First, the clay percentage is the least accurate of the determinations as it is derived from a non-basal peak whose height is greatly affected by the crystallinity of the clay lattice. Also, because a small increase in peak height represents a large increase in clay content, the positioning of the base-line can affect the clay total. Second, the determination of feldspar percentages is complicated when more than one feldspar species is present. Slow-scan runs for a number of samples indicated that up to three species of plagioclase were present. In several instances it was difficult to determine whether certain minor X-ray peaks represented genuinely discrete plagioclase compositions. Because of this the feldspars were designated plagioclase A (28.0° 2θ), plagioclase B (27.75° 2θ), and plagioclase C (28.5° 2θ). In general the relative abundances of these plagioclase species were as follows:

plagioclase A $\gg$ plagioclase B $>$ plagioclase C.

However, in volcanogenic sediment samples plagioclase A is subequal in abundance to plagioclase B. In the Mount Egmont/Palmerston North area feldspars from Quaternary andesitic ashes show peaks at 27.8° 2θ, 28.0° 2θ and 28.7° 2θ, whereas the feldspar from "greywackes" has only a single peak at 28.0° 2θ corresponding to a more acid feldspar (Symes and Wells, 1973). To aid in the discrimination of these multiple plagioclase peaks, together with the potash feldspar peak (27.5° 2θ), all samples were scanned slowly ($\frac{1}{2}^{\circ}$ 2θ per minute) through the 26° to 29° 2θ region and the positions of the feldspars transferred to the quantitative scan. The total feldspar abundance was considered to be equal to the sum of the individual plagioclase and potash feldspar percentages.

Because of the varying degrees of reliability of the method in establishing modal composition the sum total of minerals was not recalculated to 100%. Moreover, other minerals (e.g. analcite, α-cristobalite, and ferromagnesian minerals) may be present in sufficient quantity to significantly lower the total count. If any one other mineral was present in sufficient quantity a rough estimate of its abundance could be made by subtraction.

The presence of non-clay minerals was recorded in both bulk samples and the $< 2\mu$ size fraction of samples using the peak positions indicated in Table 2.3, but no attempt was made to quantify their presence in the clay fraction.

Mineral	d Å	h k l	Remarks
Quartz	3.34 4.26	101 100	
Plagioclase A	3.18	002	Oligoclase – Andesine
Plagioclase B	3.21	20$\bar{2}$, 040	Andesine – Labradorite
Potash feldspar	3.24	040	
Amphibole	8.45	110	Corresponds to plane of perfect cleavage
α -cristobalite	4.05	101	Broad-based peak
Analcite	3.43 5.57	400 211	

Table 2.3. XRD data for the identification of non-clay minerals present in samples. Plagioclase feldspar composition determined by optical microscopy.

CLAY MINERAL ANALYSIS

This section describes the methods of clay mineral preparation and analysis, with special reference to techniques found applicable for semiquantitative XRD analysis. The $<2\mu$ size fraction of all samples was analysed (Appendix

Tables IV.1 - IV.3). To complement this data, however, 39 samples representative of most onshore formations and some offshore locations were further size fractionated into < 1μ, < 2μ, 2-4μ, 4-8μ and 8-16u fractions. Most offshore samples were too small for particle size fractionation analysis. Because no universally accepted standard procedure exists for size fractionating clay mineral suspensions, the technique used in this study is detailed in Appendix Table V.1. The distribution of clay mineral species in the various size fractions is listed in Appendix Table V.2 and summarised in Figs. 2.4 and 2.5.

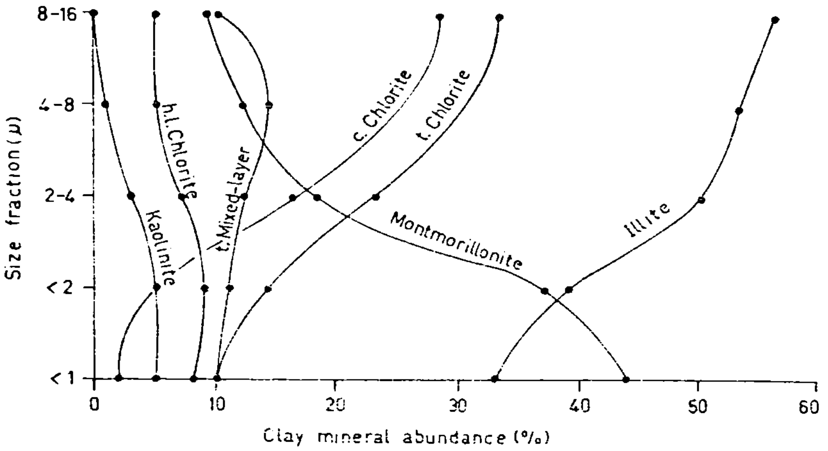


Fig. 2.4. Distribution of major clay mineral species in selected size fractions. Note that montmorillonite, heat labile (h.l.) chlorite and kaolinite are relatively more abundant in the smaller particle size fractions, whereas illite, crystalline (c) chlorite and total (t) chlorite are relatively more abundant in the larger particle size fractions.

The data indicate that the major clay mineral species are size segregated in samples and are most abundant in the following fractions: montmorillonite in < 1μ, kaolinite in < 2μ, mixed-layer clays in 4-8μ, and chlorite and illite in

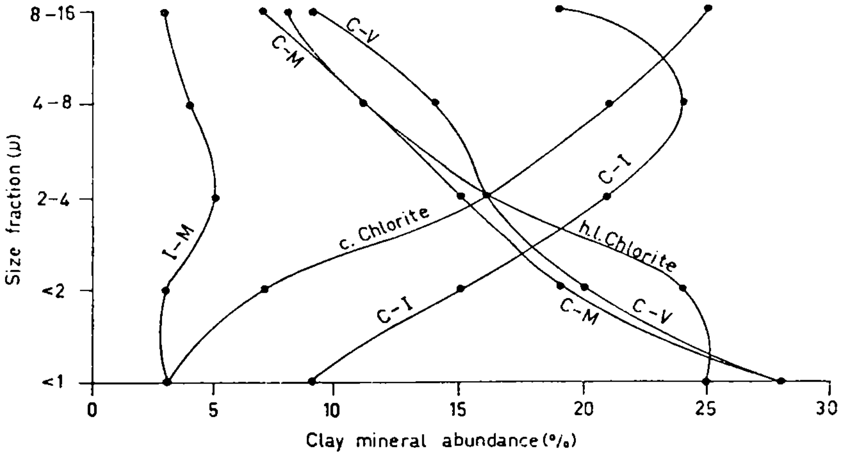


Fig. 2.5. Distribution of chlorite species and mixed-layer clay minerals in selected size fractions. Note that crystalline (c) chlorite and chlorite-illite mixed-layer clay (C-I are relatively more abundant in the larger particle size fractions, whereas heat labile (hl) chlorite, chlorite-montmorillonite mixed-layer clay (C-M) and chlorite-vermiculite mixed-layer clay (C-V) are relatively more abundant in the smaller particle size fractions.

8-16μ. Transmission electron micrographs of clay fraction separates confirm the abundance of montmorillonite in finer size fractions (cf. Pls. 3.6, 3.11 and 3.16). Mixed-layer clay and chlorite species (see pp. 45-51) are also size-segregated (Fig. 2.5). These data support findings of other workers (e.g. Grim, 1953; Gibbs, 1965) and indicate that particle size segregation will influence significantly the geologic interpretation of clay mineral data (Fig. 2.6; cf. Towe, 1974). Although the bulk of the clay minerals do occur in the <2μ fraction, the bias included in the selection of the traditional <2μ fraction is compensated for, in this study, by analysis of other size fractions.

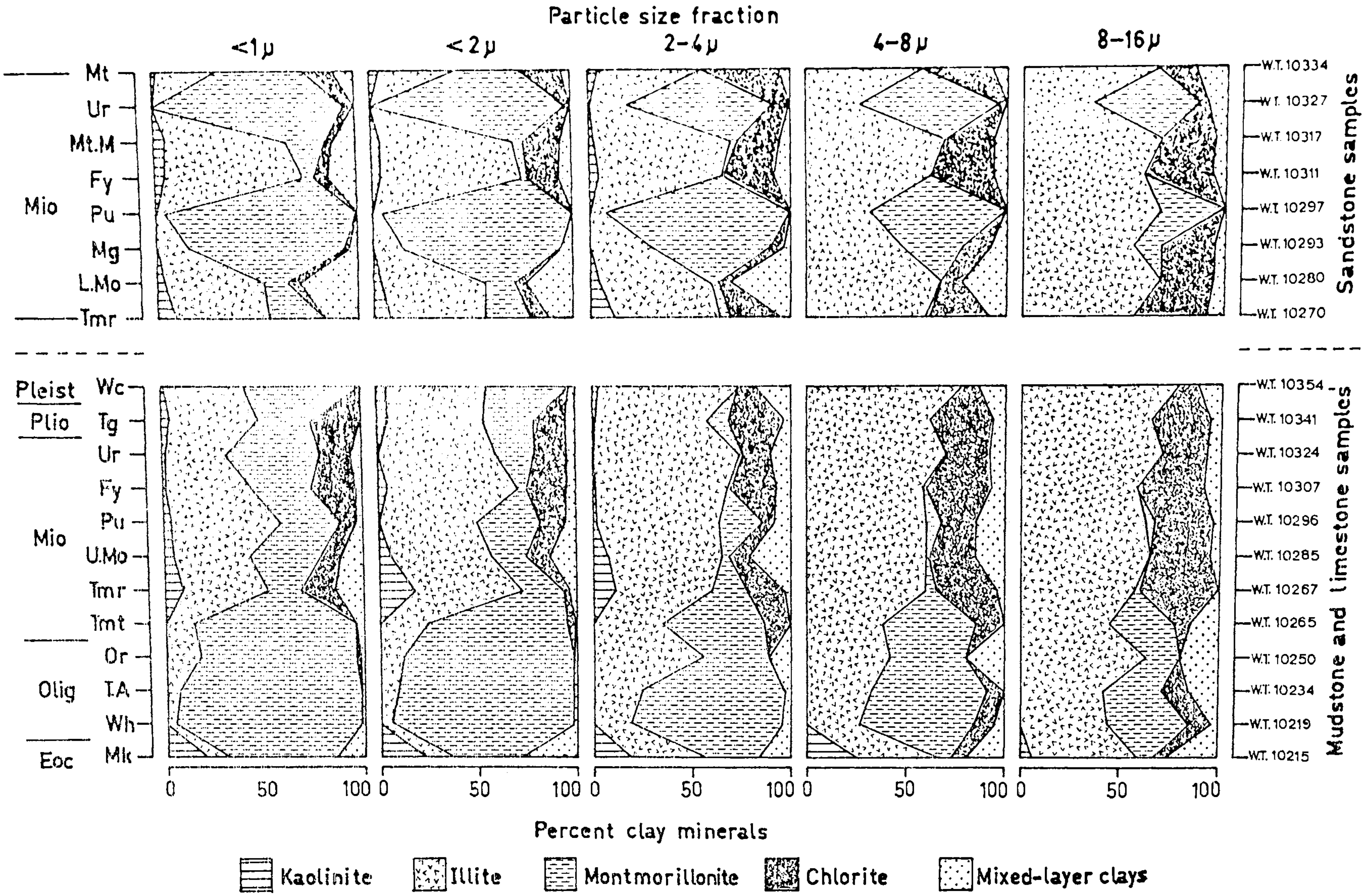


Fig. 2.6. Diagram illustrating the influence of particle size segregation on the geologic interpretation of clay mineral distributions in sediments. Abbreviations of formation names defined in Table 1.1. Note the tendency for montmorillonite to dominate the smaller particle size fractions and for illite and chlorite to dominate the larger particle size fractions.

Sample Preparation and Analysis

Methods of sample preparation have been described earlier (Table 2.1). The several different methods available for mounting clay minerals for XRD analyses have been critically reviewed by Gibbs (1965). Those used in the present study are described briefly below:

(1) Dropper-on-glass slide (DOGS) method: Most clay fractions in this study were mounted using this technique. An aliquot of clay suspension was shaken and an oriented clay mount prepared by transferring with an eye-dropper some 2-4 ml of clay suspension on to each of three microscope slides (25 x 32 mm). The specimens were evaporated to dryness under atmospheric conditions in a dust-free container. Although the DOGS method has been criticised (Gibbs, R.J., 1965, 1968), the reproducibility of the technique is good, and the patterns of clay mineral distributions are correct and produce a coherent picture (Stokke and Carson, 1973). No attempt was made to standardise the amount of solid placed on each slide because the optimum clay thickness required to produce even, non-curling surfaces varied from sample to sample. Stokke and Carson (1973) have shown that the relative intensities of major clay mineral XRD peaks can vary with the quantity of sample mounted. In this study the clay density in sample-mounts ranges from 1 to 3 mg/cm^2 which, on the basis of Stokke and Carsons' (1973) data, suggests that the maximum variation in clay mineral abundances would be 8% for montmorillonite (17A^o), 6% for illite (10A^o), and 2% for kaolinite (7A^o).

(2) Smear-on-glass slide (SOGS) method: The extremely

low settling velocity of montmorillonite meant that it was necessary to concentrate the clay suspensions in a super-centrifuge. After decanting the supernatant liquid, the remaining paste was smeared, with one stroke, onto a microscope slide (25 x 32 mm) using a flexible plastic spatula.

(3) Powder-press method: A number of <2μ particle-size suspensions were dried at 40°C, crushed to a fine powder, and mounted in Philips XRD powder sample holders using the back-filling technique. These unoriented powder samples were analysed to determine the position of the d (060) reflections of chlorite and illite.

XRD analysis of untreated sample mounts indicated that several further treatments were necessary for the positive identification of clay mineral species. These are described below:

(1) The untreated slide was glycolated using the vapour pressure method (Brunton, 1955) to determine the presence of montmorillonite. A desiccator with about 2 cm of ethylene glycol in the bottom was found to be a convenient, portable container for this purpose. Specimens were placed in this container for 12 hours, heated at 60°C for a further 1 hour, and X-rayed after cooling.

(2) The glycolated slide was heated at 500°C for 1 hour to determine the nature of the chlorite minerals present (see pp. 45-51).

(3) Clay mineral suspensions were boiled in 10% HCl for 5 minutes to distinguish, where necessary, between kaolinite and chlorite (Brindley, 1972).

(4) Selected specimens were saturated in 1NKOH for

15 hours, then heated in 1NKOH at 90°C for 1 hour (Weaver, 1958c). The degree of contraction of montmorillonite (001) is believed to indicate the nature of the parent material of the montmorillonite.

Gibbs R.J. (1965, 1968) cited the SOGS method as one producing clay mineral percentages closest to "true" values. He suggested that mounting methods involving settling in aqueous solution (e.g. DOGS method) produce large errors as a result of the surface segregation of the smallest clay particles (mainly montmorillonite) caused by differential settling of various minerals. Forty-eight samples were prepared using each of the DOGS and SOGS methods so as to compare the efficiency of, and the relationship between, the two techniques (Appendix, Table VI.1). The same qualitative results were obtained using both methods but compared to the SOGS mounts peak resolution in some DOGS preparations was low if the sample contained very fine-grained amorphous material e.g. iron oxides. Presumably these fines become concentrated on the surface of the mount and partly screen the clay minerals beneath from the X-rays.

Visual inspection of the data (Figs. 2.7 A-C) suggests that the two methods give comparable results. Statistical analysis was carried out on the data to compare the two techniques (Appendix VI). The linear correlation coefficient values of > 0.9 indicate that for montmorillonite, illite and chlorite + kaolinite + mixed-layer clays, the DOGS versus SOGS relationships are strongly linear. A one way analysis of variance shows that montmorillonite, illite and chlorite + kaolinite + mixed-layer clay mineral abundance calculated

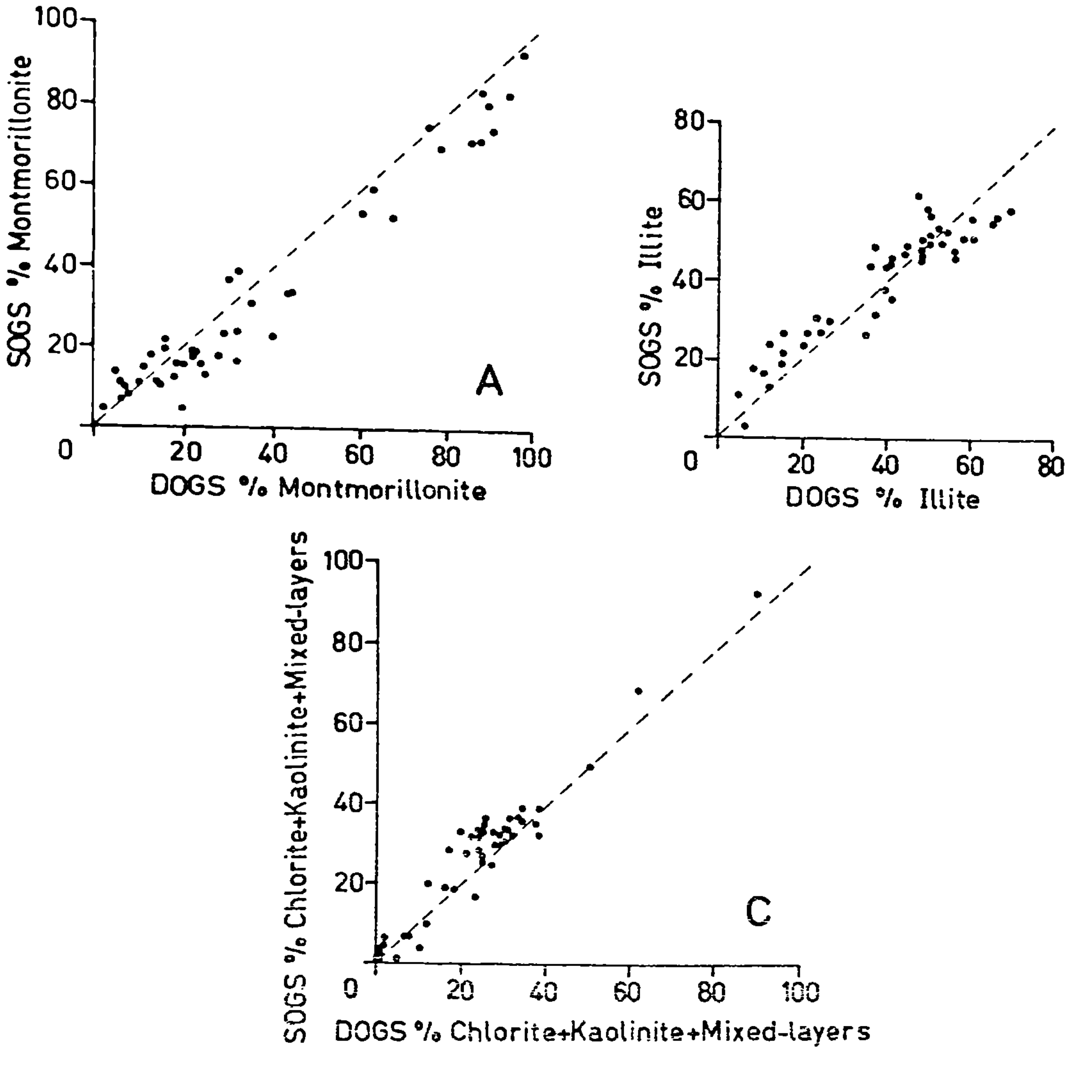


Fig. 2.7. Comparison of the quantitative distributions of montmorillonite (A), illite (B), and kaolinite + chlorite (C) in the <2μ size fraction from 48 samples using the dropper-on-glass slide (DOGS) and the smear-on-glass slide (SOGS) mounting techniques. The dashed lines are drawn to show a 1:1 relationship between DOGS and SOGS data.

from mineral mounts prepared by both the DOGS and SOGS techniques differ significantly at the 99% confidence level. Clay mineral percentages determined using the two methods are compared in Figures 2.7 A–C. The scatter of values about the line are, in part, a function of inconsistencies inherent in slide preparation techniques that can cause variations in clay mineral abundance up to ± 10% (Stokke and Carson, 1973). The distribution of data points predominantly above the line in the case of montmorillonite and predominantly below the line in the case of chlorite + kaolinite + mixed-layer clays, suggests that compared to the SOGS technique the DOGS technique gives montmorillonite percentages that are slightly higher and chlorite + kaolinite + mixed-layer clay values that are slightly lower. This is comparable to trends noted by other workers (e.g. Grim, 1953; Gibbs, 1965; 1968).

The differences in absolute values obtained using the DOGS as opposed to the SOGS technique alter only slightly the patterns of clay mineral distributions (Fig. 2.8) derived from either method. For this reason the less time consuming DOGS method was used for slide preparation in this study.

Semiquantitative analysis of clay minerals was carried out on the same machine and at similar machine settings to those used for modal analysis (p. 22). Samples were run from 3° to 35° 2θ and some slow scans ($\frac{1}{2}^{\circ}$ 2θ per min.) were made over the 3° to 14° 2θ region to allow more accurate determinations of clay mineral peak positions. A number of specimens were examined for regularly interstratified (superlattice) clays by scanning slowly over the 2 to 5° 2θ region with narrow slit settings ($\frac{1}{4}^{\circ}$).

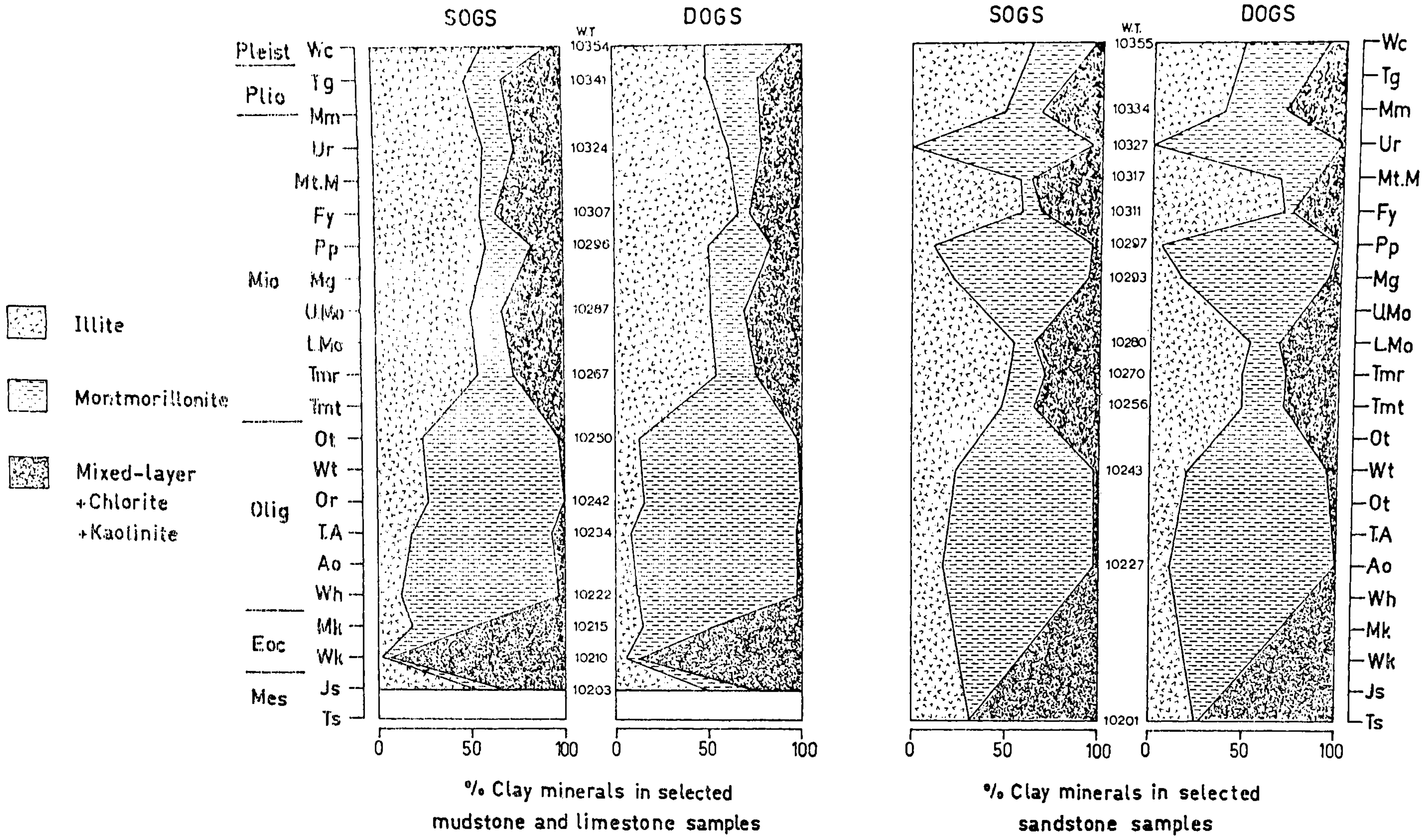


Fig. 2.8. Comparison of the trends in clay mineral distribution obtained using the dropper-on-glass slide (DOGS) and the smear-on-glass slide (SOGS) techniques of clay mineral mount preparation. Note that for each method clay mineral trends are comparable. Abbreviations of formation names defined in Table 1.1.

Qualitative Evaluation of Clay Mineral Data

The clay mineral species occurring in this study were identified by comparing their characteristic basal X-ray diffraction maxima. The positions of the (001) reflections for each mineral after various treatments are summarised in Table 2.4. Several non-clay minerals, including quartz, plagioclase, potash feldspar, α-cristobalite, amphibole and analcite, also occur in the <2µ particle size mounts, and their distributions are recorded.

Illite: Illite was identified by the strong basal (001) reflection at 9.82 Å in the untreated samples which remained unaffected after glycolation, heat and acid treatments. The degree of crystallinity of illites is indicated by the shape of the peaks, with fine-grained, poorly crystalline illites having broad peaks and highly crystalline illites (or muscovites) sharp narrow peaks. Where applicable, the crystallinity index of illite was determined using a method similar to that of Kubler (1966). In this study the crystallinity index was defined as the ratio of the peak width at half-peak height to the peak height above the background trace. This useful parameter could be determined rapidly on initial scans of samples. It is known that the chemical composition of the octahedral layer of illites determines the relative intensities of the 5 Å (002) and the 10 Å (001) peaks (Brown, 1955). Those

Table 2.4.

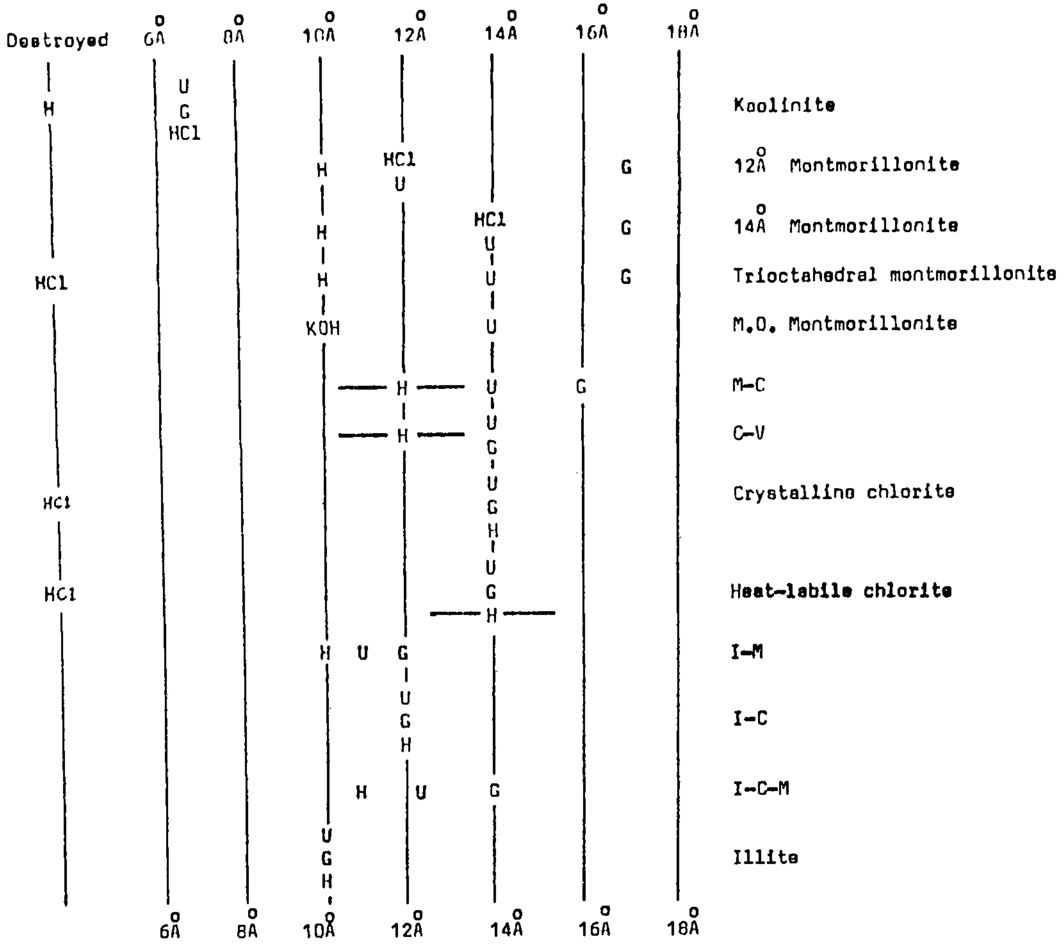
Diagram illustrating the movement of (001) X-ray diffraction peaks when clay minerals are subjected to various treatments.

U = untreated sample,
G = glycolated sample,
H = heated at 500 °C for 1 hour,
HCl = boiled for 5 min. in 10% HCl,
KOH = 1 N KOH for 15 hr followed by 1 N KOH at 90 °C for 1 hr

—— H —— = broad resultant peak

M.O. Montmorillonite refers to montmorillonite derived from the alteration of micaceous material. Mixed-layer clay minerals are designated as follows: montmorillonite-chlorite (M-C) chlorite-vermiculite (C-V), illite-montmorillonite (I-M), illite-chlorite (I-C), and illite-chlorite-montmorillonite (I-C-M).

(After Millot, 1970).



micas with octahedral layers rich in iron and magnesium (e.g. biotites) tend to have a low $\left\{\frac{(002)}{(001)}\right\}$ ratio while those with more aluminous octahedral layers have high ratios. This ratio was determined for illites from all the mudstone and sandstone samples analysed.

Montmorillonite: The mineral name, "montmorillonite" refers to both a group of clay minerals and to a member of the group (Ross and Hendricks, 1945). Minerals whose (001) peak expanded to 17 Å on glycolation were designated as montmorillonite. Before treatment the (001) peak ranges from 12 to 15 Å and after heating collapses to 9.82 Å. From the position of the (001) peak it is possible to estimate the type of exchangeable cations present: 12.4 Å - one layer of water and probably Na^{+} ; 14 to 15 Å - 2 layers of water and probably Ca^{2+} and/or Mg^{2+} (Weaver, 1958a, p.268). On this basis X-ray diffractograms indicate that most montmorillonite in both the offshore and onshore sediments contains Ca^{2+} and/or Mg^{2+} as the exchangeable cation associated with two layers of water. Some basal Te Kuiti Group montmorillonites (e.g. Mk : 10214; GM : 10216, 10217, 10218) may contain Na^{+} as the exchangeable cation associated with one layer of water. A useful parameter for describing montmorillonites is the crystallinity, $\frac{V}{p}$, as defined by Biscaye (1965, p.805), where V is the depth measurement of the "valley" on the low angle side of the peak and P the height of the peak above background. Thus a perfectly crystalline montmorillonite has a $\frac{V}{p}$ value of unity whereas less crystalline montmorillonites are represented by smaller or even negative fractions. Montmorillonites examined in this study showed a wide range of crystallinities. Many montmorillonites exhibit broad-based peaks indicative of small particle size. Electron microscopy confirmed this (cf. Plate 3.11).

Although the montmorillonites had intense 17 Å (001) reflections they were characterised by extremely weak higher order basal reflections (Table 2.5) that is typical of layer silicates with random stacking (i.e. turbostratic arrangement), parallel to the _a_ and _b_ crystallographic axes (Brindley, 1967, p.15).

001	d Å	Intensities of basal reflections for a glycolated montmorillonite
001	17.00	100
002	8.67	4
003	5.64	2
004	4.92	0.5
005	3.35	3

Table 2.5. Intensities of basal reflections for a glycolated montmorillonite in the $<2\mu$ fraction from the Whaingaroa Siltstone (Sample 10219). Note the low intensities of the higher order basal reflections.

Montmorillonites provide a good example of a highly turbostratic structure due to the weak bonding between layers and to their hydrated state and complex distribution of interlayer cations (Brindley, 1967, p.15).

The montmorillonites were further categorised according to their reaction to treatment with HCl (Table 2.4). Certain montmorillonites, particularly those of known volcanic origin, dissolved following acid treatment and these were designated as trioctahedral. General speaking, dioctahedral montmorillonites are far more resistant to acid treatment (MacEwan, 1972, p.186). Weaver (1958c) has shown that treatment with KOH (cf. Table 2.4) causes montmorillonites derived from 2:1 mica-type minerals such as muscovite and biotite to collapse from 14 Å to 10 Å because of their relatively high inherited charge. Of the 16 samples treated in this study none showed complete collapse to 10 Å, many showed partial collapse (12-13 Å) and some remained unchanged (14.2 Å) (Table 2.6). This indicates that the montmorillonite in these samples is more likely to have been derived from low-charged non-micaceous minerals such as volcanic material, mafic rocks and minerals, or chlorite (Weaver, 1958c). Two of the samples treated in this study (Table 2.6; Pu:10297; Ur:10327) contained montmorillonite that was most probably neoformed from volcanic material (cf. pp. 124-128 and p.133), supporting Weaver's (1958c) argument that montmorillonite formed from volcanic material will not collapse to 10 Å when treated with KOH.

The clay mineral montmorillonite, as defined by the structural and chemical character of its basic lattice, does not absorb potassium and collapse to become illite. Illite and muscovite can have some of their interlayer potassium removed and are then described as degraded illites (Grim, 1951) or "yawning" illites (Millot, 1970), or can

Sample Number	Sediment description	Position of (001) montmorillonite (dÅ) after KOH treatment.
B 796	Western shelf mud	12.62
10376	Kawhia Harbour mud	12.62
10351	Hawera fine sand	13.59
10358	Hawera mud	13.59
10335	Matemateaonga sandstone	12.62
10327	Urenui sandstone	14.25
10324	Urenui mudstone	12.62
10323	Mt.Messenger sandstone	12.62
10307	Ferry mudstone	13.30
10297	Purupuru sandstone	12.81
10296	Purupuru mudstone .	13.81
10293	Mangarara sandstone	13.59
10285	Taumatamaire mudstone	13.00
10265	Taumatamaire mudstone	12.80
10247	Otorohanga limestone	14.40
10223	Whaingaroa siltstone	12.99

Table 2.6. Position of montmorillonite (001) on treatment of $<2\mu$ size fraction materials with KOH after the method of Weaver (1958c).

have potassium removed to the extent that water is absorbed between the layers and the clay mineral will expand on glycolation, and thus superficially resemble montmorillonite. When such clays are placed in a potassium solution they contract to 10 Å (Brown, 1953; Kunze and Jeffries, 1953; Van der Marel, 1954; Rich and Obershain, 1955; Weaver, 1958c). Of the wide variety of montmorillonitic samples treated with KOH in this study (Table 2.6) none showed collapse to 10 Å indicating that those minerals peaking at 14 Å were indeed true montmorillonite and not degraded illites. The XRD identification of illite and montmorillonite is also supported by transmission and scanning electron microscope examination of samples. Degraded illites typically have low crystallinity. (Dunoyer de Segonzac, 1970, p.319). Thus 14 Å minerals that expand to 17 Å on glycolation, and have a high crystallinity (e.g. those 14 Å minerals identified as montmorillonite in Oligocene sediments) are unlikely to be degraded illites.

Kaolinite and Chlorite: It is difficult to distinguish between kaolinite and chlorite using XRD techniques when both are present in the same clay mineral assemblage. This is due to similar d-spacings of the kaolinite (001) and chlorite (002) at 7 Å, and the kaolinite (002) and chlorite (004) at 3.5 Å. Distinction using the (060) reflections is usually not possible because of masking by d (060) reflection of other layer silicates present. The technique described by Johns *et al.* (1954), based on the preferential destruction of chlorite basal planes other than (001) by heating to 450°C for 45 mins. was attempted, but proved unsuccessful because of differences in crystallinity between samples. Distinction between kaolinite (002) and chlorite (004) at 3.5 Å using the method of Biscaye (1964) was not applicable to samples of this study. Distinction was made finally using an acid dissolution technique (Walkden, 1972); after boiling the sample for 5 minutes in 10% HCl any peak left at 7 Å was considered to be due to kaolinite. The validity of this technique was affirmed by the complete dissolution of highly crystalline chlorites, those most resistant to acid dissolution (Ross, 1969; Brindley, 1972), in samples which from transmission electron microscopy contained no kaolinite.

Chlorite and Chlorite Mixed-Layer Clays: Chlorite is an important component of many clay mineral suites in this study and is composed of non-expandable layers with a series of (001) diffraction maxima of 14 Å periodicity. The identification of chlorite and chlorite mixed-layer clay minerals (hereafter called chlorite species) proved difficult and a suitable classification scheme, detailed below, was established only after some research.

The identification of chlorite species in this work was difficult for the following reasons:

(1) The low intensity of the 14 Å (001) chlorite peak.

(2) The masking of the (001) chlorite peak by other minerals, notably montmorillonite.

(3) The presence of heat-labile chlorite species whose (001) peaks collapse on heating at temperatures as low as 350°C and may thus be mistaken for vermiculite.

(4) The wide spectrum of positions and shapes of the (001) chlorite peak following heating at 500°C for one hour (a routine test adopted for the identification of chlorites).

All chlorites examined show pronounced (002) and (004) reflections but comparatively weak (001) and (003) reflections, indicating they are iron-rich varieties (Weaver, 1958a, p.266). Unoriented powder samples show a d (060) spacing of about 1.54 Å indicating the chlorites are trioctahedral (Brindley, 1972, p.267).

To deduce the nature of the various chlorite species present in the sediments two lines of approach were adopted.

(1) Ten representative samples (Table 2.7) were heated to temperatures ranging from 350°C to about 650°C for a period of one hour, and the shifts and changes in shapes of the peaks of the chlorite species were compared. On heating, the 14 Å (001) chlorite peak in different samples collapsed at temperatures ranging from 500°C to over 650°C, while the corresponding 7 Å (002) peaks collapsed at slightly lower temperatures (Table 2.7). Chlorite peaks from sandstones of interbedded sandstone/mudstone sequences tended

Sample Number	Sample Description	Temperature °C							
		350	400	450	500	550	600	650	650
8319	Recent sediment		●	●	X		●	●	O
8645	"		●	●	●	X	●	●	O
8818	"		●	●	●	X	●	⊙	
C363	"				●	X	●	⊙	
10311	Ferry sandstone.	●	●	X	⊙		●		
10307	Ferry mudstone		●	●	X		●	⊙	
10295	Purupuru sandstone	●	●	X	⊙		●		
10294	Purupuru mudstone		●	●	X	⊙	●		
10293	Mangarara sandstone	●		●	X		⊙		
10257	Taumatamaira mudstone	●	●	●	X	⊙	●		

Table 2.7. Temperatures at which heating for 1 hour produces collapse of the 14Å chlorite (O) and 7Å chlorite (X) reflections. (● = Range of temperatures at which samples were heated). Note that the 7Å chlorite peak collapses at a lower temperature than the 14Å chlorite peak.

to collapse at lower temperatures than those from mudstones from the same sequences.

The above data are explained by considering the mechanisms of chlorite decomposition as detailed by Brindley (1972). Differential thermal analysis and weight-loss determinations of chlorites show that two endothermic dehydration reactions followed by an exothermic recrystallation are involved in the dehydration and decomposition of chlorites. The first dehydration reaction corresponds to a loss of water from the brucite layers which produces small changes in lattice parameters but large changes in reflected intensities. The second dehydration reaction corresponds to a loss of water from the talc layers. The final reaction involves recrystallisation of new products, notably olivine $(Mg\ Fe)_2SiO_4$. The temperatures of the dehydration reactions, and the extent to which they are clearly separated, depend on the composition, crystallinity, particle size and thermal conditions (e.g. rate of heating) of the system. Clear demarcation between the two dehydration reactions is evident under heating conditions. With rapid heating, a small particle size favours a clearer separation of the two processes. Poorly crystalline minerals, such as those generally found in Recent sediments, decompose at temperatures as low as 300°C (Bradley, 1953); well crystallised minerals however decompose at around 600°C. With low iron/high magnesium chlorites the two dehydration reactions take place at temperatures of about 600°C and 800°C respectively. With iron-rich chlorites, and especially those of poor crystallinity and small particle size, the reactions take place at lower temperatures and are complicated both by the oxidation of iron and partial dehydration due to loss of hydroxyl ions.

In the present study, the collapse of certain chlorites under conditions of rapid heating at low temperatures accords with the fact that they are trioctahedral (Fe^{2+}), iron-rich chlorites of low crystallinity (IIb polytype) and small particle size ($< 2\mu$). The relative intensities of the (001) peaks of the more stable chlorites indicate they too are iron-rich and probably owe their stability to a higher degree of crystallinity. Mudstone samples from interbedded sandstone/mudstone sequences contain less readily decomposed (i.e. more stable) chlorite minerals, perhaps because the mudstones are far less permeable and therefore offer the least modified image of the original clay minerals composition. The less stable chlorites from the sandstones have probably formed through degradation by percolating formation fluids.

(2) The influence of mineral composition on the decomposition temperature of the chlorite species was further illustrated by observing the effects of heating oriented $<2\mu$ size fraction mounts at 500°C for one hour. The resulting XRD patterns (Fig. 2.9 A-C) show that this treatment causes the 14 Å (001) chlorite peak to do one or other of three things. It may decrease in height but remain similar in shape (Fig. 2.9A); it may decrease in height and widen (Fig. 2.9B); or it may be displaced towards 12 Å to produce a poorly defined plateau (Fig. 2.9C). The first curve is typical of iron-rich stable chlorite (Carrol, 1970,

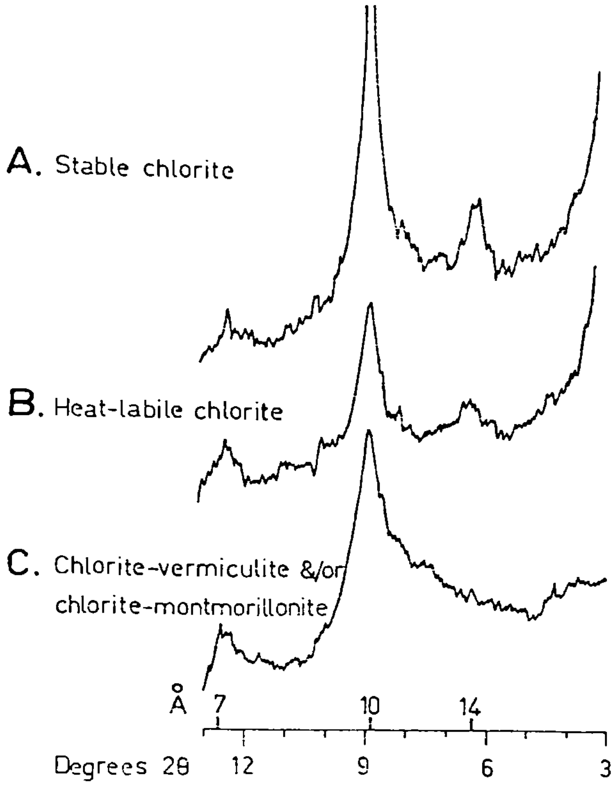


Fig. 2.9. Changes in shape and position of the (001) chlorite XRD reflection after heating <2µ sample mounts at 500°C for one hour. Stable chlorite is characterised by a sharp peak at 14Å whereas heat-labile and mixed-layer chlorites produce less intense and asymetric reflections.

p.53). The second is characteristic of heat-labile chlorite (Dunoyer de Segonzac, 1970, p.326). The third case is probably due to the presence of either disordered (randomly interstratified) mixed-layer chlorite-montmorillonite (C-M) and/or disordered chlorite-vermiculite (C-V). These two minerals can be readily differentiated as C-M imbibes ethylene glycol (Shover, 1964, p.435; Dunoyer de Segonzac, 1970, p.326). The presence of chlorite-vermiculite and/or chlorite-montmorillonite clays is verified in the glycolated sample by the asymmetry of the 7 Å (002) chlorite peak on the low-angle side and asymmetry of the 17 Å (001) montmorillonite peak on the high-angle side. The presence of chlorite-vermiculite tends to reinforce the 14 Å (001) peak relative to the 7 Å (002) peak in glycolated samples. Other XRD patterns indicate that other chlorite varieties exist as intermediates between the four major chlorite species described above. MacEwan (1972, p.186) suggested that "When dealing with an unknown clay, one must be prepared for any combination of chlorite, vermiculite and montmorillonite or gradation between them.... These different types of minerals probably grade into one another."

Two further types of chlorite mixed-layer clays have been identified. A peak at 12.3 Å unaffected by glycolation and heating indicates the presence of a randomly interstratified chlorite-illite (C-I) mixed-layer clay. The position of the peak suggests that the ratio of chlorite to illite is approximately 1:1. The presence of a three component randomly interstratified mixed-layer clay, namely illite-chlorite-montmorillonite (I-C-M), is indicated by a peak at 12.3 Å in the untreated sample which shifts to 14.2 Å on glycolation and forms a broad peak centred at 11.5 Å after heating.

In summary, the following six varieties or species of chlorite are present in the samples studied: crystalline stable iron-rich chlorite; heat-labile iron-rich chlorite; disordered mixed-layer chlorite-vermiculite; disordered mixed-layer chlorite-montmorillonite; chlorite-illite mixed-layer; and illite-chlorite-montmorillonite mixed-layer.

Mixed-Layer Clays: A peak at 10.65 Å, which increased to 11.05 Å on glycolation and collapsed to 9.82 Å on heating, was attributed to the (001)/(001) basal reflection of a randomly interstratified illite - montmorillonite (I-M) mixed-layer clay. Using the method of Weaver (1956, p.206) it was estimated that this mixed-layer mineral contained approximately 20% expandable layers. The position of the (001) "montmorillonite" peak in some Lower Tertiary samples suggests that the montmorillonite present may, in fact, contain a few percent of non-expandable layers and is not always "pure" montmorillonite.

In some samples poor resolution of clay mineral peaks did not allow positive identification of individual mixed-layer clay minerals and it was commonly convenient to label the mixed-layer clay peak in the 12 Å to 13 Å region as undifferentiated mixed-layer clay.

Chlorite mixed-layer clay minerals were discussed earlier (pp.45-50).

Semi-quantitative Evaluation of Data

Because there are many variables affecting the quantitative evaluation of clay mineral species in sediments there is some doubt as to whether true quantification will ever be obtained. One of the many variables concerns the methods involved in calculating the relative amounts of clay minerals present in samples. The major methods currently in use have been reviewed by Pierce and Siegel (1969, p.192) who gave preference to no one single calculation technique since "...each is as well founded as any other in a true 'quantitative' sense". In this study calculations of relative amounts of clay minerals were based on peak area measurements made with a variable compensating planimeter, the peak area being that area bounded by the peak and a free-hand drawn base-line constructed by the method of Porrenga (1966). The technique used was a modification of that presented by Johns _et al_. (1954). Results are considered to be semiquantitative only, but probably accurate to within ± 10% of the true value. The calculation methods described here are not necessarily applicable to all clay sediments and the following points should be emphasised:

(1) The reported clay minerals are considered to comprise 100% of the sample, whereas in some samples there are other minerals present as well as amorphous material.

(2) The refracting ability of clay minerals of the same species, which is dependent on composition and degree of crystallinity, is considered to be constant.

(3) The weighting corrections (see next section) selected for individual clay mineral species are assumed to be valid.

Calculation Method: In establishing the relative abundance of each clay mineral species in samples all components were compared to the illite phase, after first weighting the different minerals in such a manner as to compensate for their different efficiencies in scattering along the (001) row line. The steps used for the semi-quantitative analysis of the clay minerals are listed below and a worked example follows:

(1) On the XRD pattern of the untreated sample the reflected intensities (peak areas) of the 3.3 Å (illite)

and the 3.5 Å (kaolinite and chlorite) peaks were compared directly (i.e. $\frac{3.5 \text{ Å area}}{3.3 \text{ Å area}}$). In this study montmorillonite samples gave extremely weak higher order reflections and the montmorillonite component in the 3.5 Å peak could usually be neglected;

(2) On the XRD pattern of the glycolated sample peak areas at 17 Å (montmorillonite), 10 Å (illite), 11 Å (illite-montmorillonite), 12-13 Å (undifferentiated mixed-layer clay) and 14 Å (illite-chlorite-montmorillonite mixed-layer clay) were measured. To correct for low angle polarisation, the peak areas were divided as follows: 17 Å by 4, and 11 Å, 12-13 Å, and 14 Å by 2. The corrected peak areas were then divided by the 10 Å peak area, summed along with the value for $\frac{3.5 \text{ Å area}}{3.3 \text{ Å area}}$ obtained in (1), and recalculated to 100%;

(3) On the diffraction pattern of the HCl treated sample the 10 Å (illite) and 7 Å (kaolinite) peaks were measured, the 7 Å peak area divided by a factor of 2, and compared to the area of the 10 Å peak. This ratio is an expression of the proportion of kaolinite in the chlorite plus kaolinite % value calculated in the last step of (2). The kaolinite % is then calculated and the chlorite % obtained by subtraction;

(4) By noting the shape of the 14 Å peak in the diffractogram of the heated (500°C) sample (cf. Fig. 2.9) the total chlorite % can be apportioned between crystalline chlorite, heat-labile chlorite, and C-V or C-M; the mixed-layer chlorite species were distinguished using the glycolated diffractogram;

(5) Total mixed-layer % is obtained by summing individual mixed-layer percentages.

The following example of a semiquantitative calculation is for the clay fraction of a hypothetical mudstone composed of kaolinite (K), illite(I), montmorillonite (M), chlorite (C), and chlorite-illite (C-I) mixed-layer clays. Symbols U, G, H and HCl are defined in Table 2.4. The symbol $^{a}10\ \mathring{A}$ (G) refers to the area of the 10 Å peak on the glycolated pattern; other peak areas are denoted by similar notations.

$$^{a}3.3\ \mathring{A}\ (U) = 1.8\ cm^2 \qquad ^{a}17\ \mathring{A}\ (G) = 7.4\ cm^2$$
$$^{a}3.5\ \mathring{A}\ (U) = 1.2\ cm^2 \qquad ^{a}12\ \mathring{A}\ (G) = 0.4\ cm^2$$
$$^{a}10\ \mathring{A}\ (G) = 4.2\ cm^2 \qquad ^{a}10\ \mathring{A}\ (HCl) = 3.9\ cm^2$$
$$^{a}\ 7\ \mathring{A}\ (HCl) = 0.4\ cm^2$$

Clay mineral abundance recalculated to 100%

$$I = \frac{^{a}10\ \mathring{A}\ (G)}{^{a}10\ \mathring{A}\ (G)} = \frac{4.2}{4.2} = 1.00 = 39$$

$$M = \frac{^{a}17\ \mathring{A}\ (G)}{^{a}10\ \mathring{A}\ (G)} = \frac{7.4}{4.2 \times 4} = 0.84 = 32$$

$$C + K = \frac{^{a}3.5\ \mathring{A}\ (U)}{^{a}3.3\ \mathring{A}\ (U)} = \frac{1.2}{1.8} = 0.66 = 25$$

$$C\text{-}I = \frac{^{a}12\ \mathring{A}\ (G)}{^{a}10\ \mathring{A}\ (G)} = \frac{0.4}{4.2} = 0.09 = 4$$

$$\Sigma\ \overline{2.59} \qquad \overline{100}$$

$$K = \frac{^{a}7\ \mathring{A}\ (HCl)}{^{a}10\mathring{A}\ (HCl)} \times \frac{\text{illite}}{2}\% = \frac{0.9}{4.4} \times \frac{39}{2} = 3\%$$

$$C = (C + K)\% - K\% = 25 - 3 = 22\%$$

and from 14 Å (H)

C = crystalline chlorite.

Hence the clay mineral composition of the sample is = kaolinite 3%; illite 39%; montmorillonite 32%; chlorite 22% and chlorite-illite mixed-layer clay 4%.

2.3.2B TRANSMISSION ELECTRON MICROSCOPY

Transmission electron microscopy (TEM) was used to examine morphologic characteristics of clay size material to help confirm clay mineral identifications made using XRD, to identify halloysite (XRD identification of dehydrated halloysite is difficult in the presence of illite), and to search samples for fine-grained amorphous substances (e.g. allophane).

Clay suspensions prepared from 32 sediment samples (Appendix Table VII.1) were treated with 6% H_2O_2 to remove organic matter, washed in a centrifuge, dispersed using a solution of calgon (1 g/l) and distilled water, and then diluted to 0.1% concentration. An aliquot of this suspension was placed on a Formvar coated copper grid using a squeeze bulb and was allowed to air dry. An attempt was made to freeze-dry several samples using liquid nitrogen but this method resulted in the tearing of the Formvar specimen support. The prepared suspensions were shadowed with platinum from carbon electrodes at an incident angle of 45^o before examination.

Mineral identification was made by comparing the electron micrographs with those displayed in other works (e.g. Fieldes and Williamson, 1955; Beutelspacher and Van der Marel, 1968; Oertel, 1973) and by examining the electron micrographs in the light of XRD data from the same sample.

No attempt was made to quantify clay mineral abundances as the identification of illites, chlorites and mixed-layer clays in multicomponent mixtures is difficult and the extreme subsampling that is a necessary part of sample preparation casts doubt on the qualitative representativeness of the subsample.

Most electron micrographs showed subhedral and anhedral mineral grains. This is an important characteristic as in general such crystal shapes are common to detrital clay minerals and reflect crystal degradation during transport (Beutelspacher and Van der Marel, 1968; Sarkisyan, 1972, p.8.). "Wavy" patterns on some of the larger clay minerals (e.g. Pl.3.26 and 3.28) are produced by irregularities in the lattice and by the bending of plates during weathering (Beutelspacher and Van der Marel, 1968, p.116), or represent interference figures caused by stress as a result of weathering (cf. Beutelspacher and Van der Marel, 1968, p.144 Fig. 137). Authigenic clay minerals generally show euhedral crystal form and numerous examples of these can be found in the literature (e.g. Beutelspacher and Van der Marel, 1968; Millot, 1972, pp.220, 222 and 345-345; Sarkisyan, 1972). Examples of euhedral crystals from sediments of this study can be seen in Pls.3.4, 3.12, 3.13, 3.21 and 4.16.

In the $<2\mu$ size fraction of several samples (e.g. 10227, Pl.3.11; 10236, Pl.3.12; 10219, Pl.3.13; 10296, Pl.3.21; 10359, Pl.3.34) small (0.03 μ diameter) rounded "blebs" are distinguished. The size and shape of the "blebs" does not compare favourably to known clay mineral morphology.

XRD analysis of these samples indicates that they consistently contain small quantities of α- cristobalite and it is possible that the "blebs" represent the α- cristobalite component in these samples. Clay-size cristobalite has been described elsewhere in the literature (e.g. Wermund and Moila, 1966; Reynolds, 1970) but "blebs" similar in shape and size to those found in this study were only seen figured in Beutelspacher and Van der Marel (1968, p.218, Fig. 209).

Also in the <2μ size fraction of some samples (e.g. 10236, Pl.3.12; 10219, Pl.3.13), lath-shaped aggregates of montmorillonite occur. Such aggregates are typical of neoformed montmorillonite (Roy and Sand, 1956; Kotelnikov, 1965). They are generally rare, perhaps because mechanical and chemical dispersion procedures are likely to damage or destroy such delicate structures.

2.2.2C SCANNING ELECTRON MICROSCOPY

Scanning electron microscopy (SEM) was used to examine grain fabrics and help determine sample composition. SEM is ideal for fabric studies of fine-grained sediments as it has the ability to produce good resolution at high magnification with large depth of focus. This technique was particularly suitable for this study as most sediments are very fine-grained and conventional point-counting techniques would have been both inadequate and inaccurate. Moreover many samples are unconsolidated or friable and contained swelling clay minerals which would have made thin section preparation both difficult and time consuming. SEM was used to examine 28 samples (Appendix Table VII.1).

Freshly fractured sediment chips of about 4mm size were wrapped in tissue paper and placed in air-tight 30 ml glass "snap-cap" vials along with some silica gel some 48 hours before analysis. After drying, the samples were mounted in groups of 10 in a circular pattern on 2 cm diameter brass sample stubs using clear fingernail polish as the mounting medium. Seston collected on millipore filter medium was dried and mounted on brass sample stubs using double-sided sellotape as the mounting medium. The samples were coated with gold in a rotary shadower and stored in a desiccator prior to analysis.

Mineral identification was made by comparing the electron micrographs to those displayed in other works (e.g. O'Brien, 1970; Hughes and Bohor, 1971; Eswaran, 1972; Sarkisyan, 1971; Sarkisyan, 1972; Wilson and Pittman, 1977) and by examining the electron micrographs in the light of XRD and TEM data from the same sample.

The sediments showed a variety of grain fabrics but unorientated fabrics predominate (e.g. Pls.3.10, 3.17 and 3.30). Such random grain fabrics are characteristic of many marine mudstones and while they may represent a primary depositional fabric they may also result from biogenic reworking of the sediment (cf. Byers, 1974). The only well orientated grain fabric was that from a kaolinite-rich Waikato Coal Measures sample (10210, Pl.3.5) in which the grain fabric parallels the plane of the bedding.

Scanning electron microscopy does illustrate the considerable differences in grain fabric that may exist on a micro-scale between what are otherwise in hand specimen similar mudstone samples (cf. Wh: 10222, Pl.3.10 to Tg: 10341, Pl.3.32). Purpuru Volcanic Sandstones (e.g. 10297, Pls.3.18 and 3.19) exhibit mammillary α-cristobalite whose relationship to feldspars and montmorillonite suggests it formed subsequent to these minerals and that it is, in fact, diagenetic. Two samples (Mk: 10215, Pl.3.8; Pu: 10296, Pl.3.20) contained concretionary bodies of montmorillonitic clay that possibly represent aggregates of reworked soil material.

2.3.2D X-RAY FLUORESCENCE SPECTROMETRY

Using X-ray fluorescence (XRF) spectrometry the concentrations of selected elements were determined for the insoluble residues of all sediment samples and for a number of clay fractions separated from these samples Appendix Tables VIII.1 - VIII.4. Advantages of XRF spectrometry include:

(1) Qualitative scans of samples are simple and rapid;

(2) After initial "setting-up" is completed the quantitative analysis of samples is also rapid;

(3) The analytical process is non-destructive and samples can be used for further analysis; and

(4) Only small samples are required.

The data shows no significant variations in bulk rock chemistry although they do confirm trends in composition that were predicted from preceding mineralogic analyses. For this reason the techniques, results and interpretation of XRF analyses are included only in Appendix VIII.

2.3.2E INFRARED ABSORPTION SPECTROMETRY

Infrared (IR) absorption spectrometry has been used for the identification of minerals whose characteristic absorption bands lie in the 300-4000 cm^{-1} (infrared) region.

To prepare samples for analysis the $<2\mu$ size fractions were dried at $60^{o}C$, finely ground, mixed with powdered KBr to a concentration of 0.1-1%, and pressed into discs under a pressure of 8 tons. The discs were scanned immediately using a Shimadzu double beam spectrophotometer with KBr optics in the 400-4000 cm^{-1} range at the following recording conditions: attenuation 75%, scale expansion 1:1, scanning speed 3 cm^{-1} per min.

The identification of discrete clay minerals using IR spectra proved difficult because of the presence of interstratified clays and the multicomponent clay mineralogy of samples. The results confirmed data on the nature of the clay minerals obtained using other techniques, but rarely provided additional information. IR absorption spectra of the $<2\mu$ fraction of 4 sediment samples are compared to their clay mineral abundance determined by XRD in Fig. 2.10.

2.4 SUMMARY AND CONCLUSIONS

Samples for bulk and clay mineral analyses were collected from onshore exposures of Mesozoic to Recent sediments in the Hamilton, Taranaki and Wanganui districts (Fig.2.1), from Recent surficial sediments on the western continental shelf (Fig.2.2), from harbour and river mouths bordering the western shelf, and of western shelf seston (Figs. 2.1 and 2.2). Sample analysis procedures were determined that were simple, rapid, and applicable to this

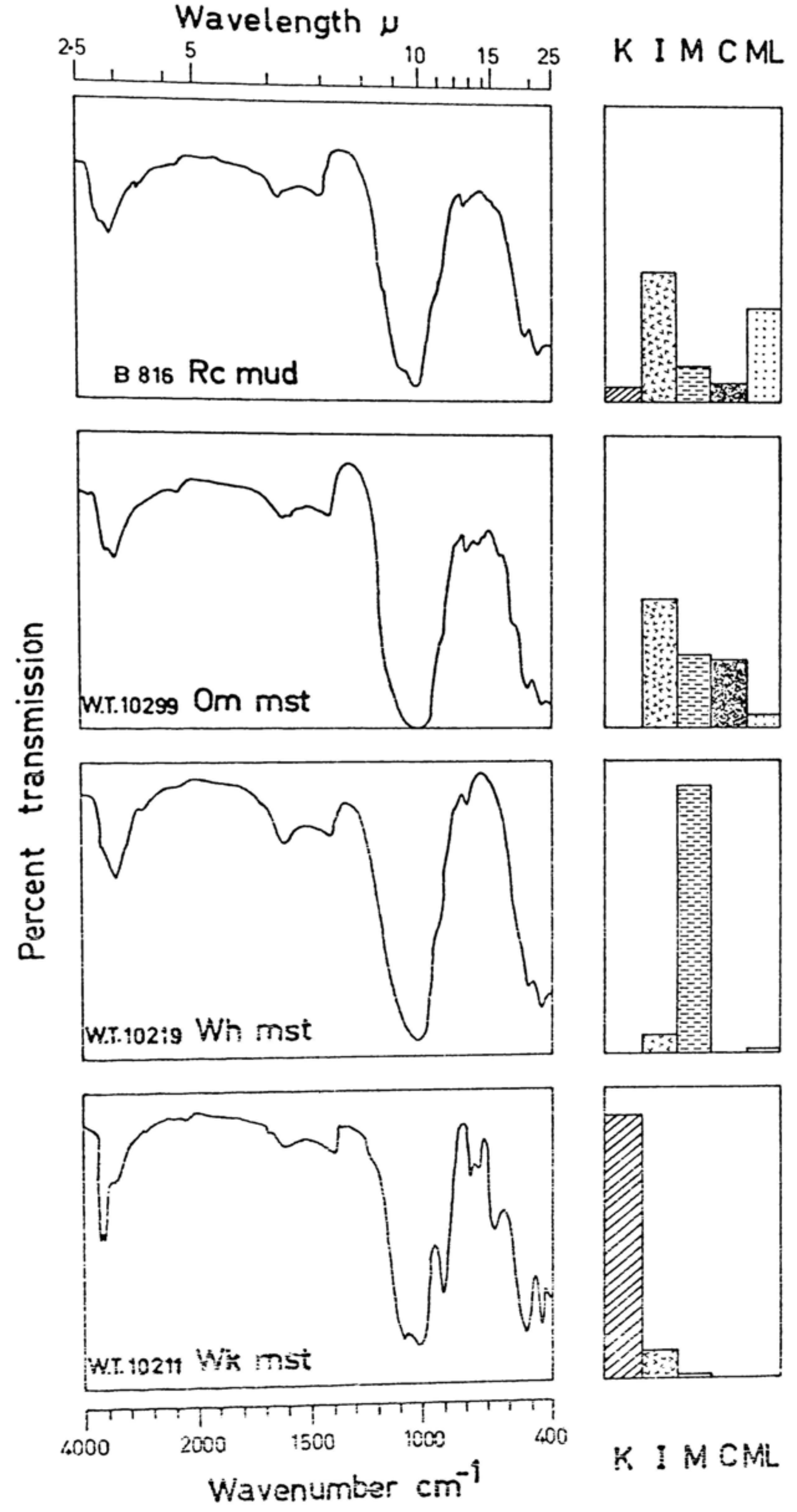


Fig. 2.10. Infrared absorption spectra of the $<2\mu$ size fraction of a Waikato Coal Measures mudstone (Wk), a Whaingaroa mudstone (Wh), an Omoao mudstone (Om), and a western shelf mud (Rc). Histograms indicate the relative abundance of kaolinite (K), illite (I), montmorillonite (M), chlorite (C) and mixed-layer clay minerals (ML) in the clay fractions as determined by XRD analysis.

wide variety of sediment types (Table 2.1).

X-ray diffraction proved to be the most useful and rapid method of quantitative bulk mineral analysis for quartz, plagioclase feldspar, potash feldspar and clay minerals. Calcite and aragonite contents were determined by acid digestion.

Clay mineral identification was made following XRD analysis of untreated samples (Table 2.4). Semiquantitative calculations of the amounts of clay minerals in samples were made using a modification of the method of Johns et al. (1954). The crystallinity index was determined for the illite, chlorite and montmorillonite component of samples where applicable. Analysis was carried out on the <2μ size fraction of all samples, and on the <1μ, <2μ, 2-4μ, 4-8μ and 8-16μ size fractions of 39 selected samples to compensate for any bias inherent in the selection of the traditional <2μ fraction. The predominance of illite and chlorite in the larger size fractions compared to montmorillonite and kaolinite in the smaller size fractions (Fig. 2.4) can significantly alter geologic interpretation of clay mineral distributions in sediments (Fig. 2.6). Clay minerals were mounted for XRD analysis using the dropper-on-glass slide (DOGS) technique. Comparison of this mounting method with the smear-on-glass slide (SOGS) technique produced the same qualitative results but the peak resolution in some DOGS preparations was poorer and gave montmorillonite percentages that were slightly higher and chlorite + kaolinite + mixed-layer clays values that were slightly lower than SOGS values (Fig. 2.7). These differences are less than those predicted by Gibbs, R.J. (1965; 1968) and show that the patterns and interpretation of the clay mineral distributions derived from either method (Fig. 2.8) are similar.

The montmorillonite minerals investigated are mostly dioctahedral, show a turbostratic stacking arrangement, contain two layers of water and have Ca^{2+} and/or Mg^{2+} as the exchangeable cation. Montmorillonites were further categorised according to their reaction to treatment with HCl. Trioctahedral montmorillonites of "known" volcanic origin dissolved following HCl treatment. While it is accepted that trioctahedral montmorillonites are acid-labile (MacEwan, 1972, p.186), the trioctahedral acid-labile montmorillonites in this study were all of volcanic origin. Hence for the onshore sediments investigated, at least, the HCl treatment (the same as that used in the identification of kaolinite) provides a simple test for volcanogenic montmorillonite.

The identification of chlorite and chlorite mixed-layer clays proved difficult, but heating tests indicated the presence of a spectrum of chlorite species including crystalline stable iron-rich chlorite, heat-labile iron-rich chlorite, disordered mixed-layer chlorite-vermiculite, disordered mixed-layer chlorite-montmorillonite, mixed-layer chlorite-illite, and mixed-layer illite-chlorite-montmorillonite (Figs. 2.5 and 2.9). Criteria for recognising this important mineral sequence are described.

Transmission electron microscopy provides data on clay particle morphology. The majority of clay minerals are anhedral and probably detrital in origin but aggregates of euhedral montmorillonite laths and fine grained α-cristobalite occur locally.

Scanning electron microscopy gives information on grain fabrics. Most samples exhibited unorientated grain fabrics although mudstones from the Waikato Coal Measures show a well orientated grain fabric.

X-ray fluorescence spectrometry was used to determine the concentrations of selected elements (Al, Si, K, Ca, Ti, Mn and Fe) in the insoluble residues of all samples and for a number of clay fractions. Infrared absorption spectrometry confirmed the nature of clay minerals provided by other techniques, but as the identification of discrete clay minerals in the presence of interstratified clays proved difficult the technique was not pursued.

CHAPTER 3

CLAY PETROLOGY OF MESOZOIC TO RECENT ONSHORE SEDIMENTS

3.1 AREA OF STUDY AND GEOLOGIC BACKGROUND

The onshore study area includes the Hamilton, Taranaki and Wanganui Districts and covers some 25,000 km^2 (Fig. 1.1). Structural basins of Cenozoic sedimentation included within this area are the Waikato, North Wanganui, South Wanganui and Taranaki Basins (Fig. 3.1).

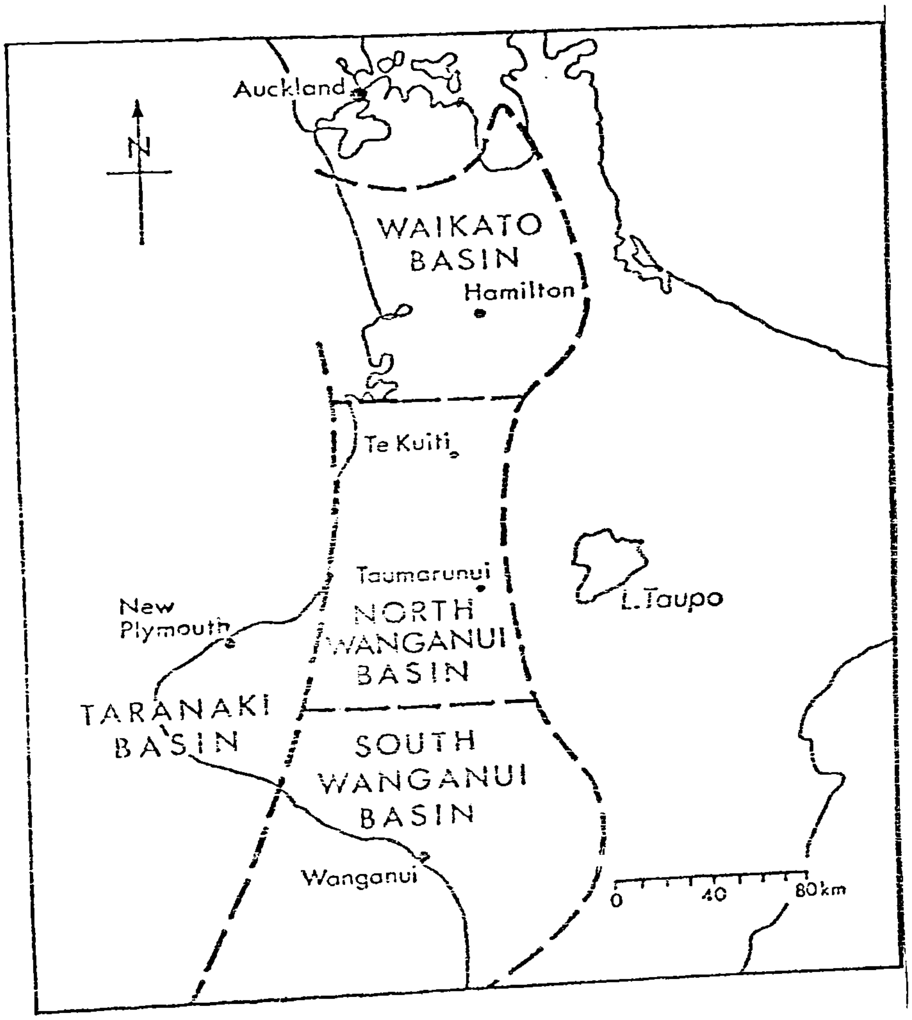


Fig. 3.1. Structural sedimentary basins in the western North Island (after Cope and Reed, 1967; Katz, 1968; Nelson, 1973).

3.1.1 STRATIGRAPHY

The stratigraphy, distribution and general lithologic characteristics of sediments exposed in the study area are summarised in Table 3.1 and Fig. 1.3, and in Tables 3.2 to 3.4. The reader is referred to Pls. 1.1 to 1.24 for photographic illustrations of selected sedimentary units.

<table>
<tr><th></th><th>Hamilton District</th><th>Taranaki District</th><th>Wanganui District</th></tr>
<tr><td rowspan="4">Quaternary</td><td colspan="3">Holocene</td></tr>
<tr><td>Hawera Stage (H)</td><td rowspan="3">▨</td><td>Hawera Stage (H)</td></tr>
<tr><td>Castlecliffian Stage(Wc)</td><td>Castlecliffian Stage (Wc)</td></tr>
<tr><td>Nukumaruan Stage (Wn)</td><td>Nukumaruan Stage (Wn)</td></tr>
<tr><td>Pliocene</td><td rowspan="4">▨</td><td colspan="2">Tangahoe Mudstone</td></tr>
<tr><td rowspan="6">Miocene</td><td>Matemateaonga Sandstone</td><td rowspan="8">▨</td></tr>
<tr><td>Urenui Siltstone</td></tr>
<tr><td>Mt. Messenger Sandstone</td></tr>
<tr><td colspan="2">Mohakatino Group</td></tr>
<tr><td colspan="2">Mokau Group</td></tr>
<tr><td colspan="2">Mahoenui Group</td></tr>
<tr><td>Oligocene</td><td colspan="2">Te Kuiti Group</td></tr>
<tr><td>Mesozoic</td><td colspan="2">Mesozoic</td></tr>
</table>

Table 3.1. Generalised correlation diagram of the major sedimentary units exposed in the study area. Hatched areas generally indicate that sediments of that age were either not deposited or have been eroded.

The mineralogy of Recent sediments, including soils, was not investigated in any detail in this study. The characteristics of the soil and tephra material in the study area are summarised on pp. 10-12.

	N.Z. Stages			Thickness (in metres)	
Lr–Mid. Miocene	Sl		Mohakatino Formation : Andesitic tuffaceous sandstones and siltstones (confined to south-western area).	250	Average thickness
	Sc | Pl	Mokau Group	Mokau Sandstones : Massive and bedded sandstones, poorly fossiliferous and calcareous. Massive sandstones frequently leached and iron stained.	260	
	Lw–Pl		Mahoenui Formation : Blue-grey massive calcareous siltstone with thin sandy limestone beds overlain by sandstone and rare calcareous siltstone.	215	
Oligocene	Lw	Te Kuiti Group	Otorohanga Limestone : Cream coloured, flaggy, crystalline limestone.	43	Thickness in Type Section area
			Waitomo Sandstone : Massive, calcareous slightly glauconitic fine sandstone, commonly leached and weathered.	18	
			Orahiri Limestone : Upper and lower flaggy crystalline limestones separated by a layer containing oyster bands and calcareous glauconitic sandstone seams.	15	
			Te Akatea Siltstone : Brittle light chocolate brown fine calcareous siltstone. Locally contains sandstone bands, cherty layers and a basal limestone.	20	
			Aotea Sandstone : Grey calcareous fine to medium sandstone, locally glauconitic and contains the Waitetuna Limestone member.	15	
			Whaingaroa Siltstone : Medium blue-grey calcareous siltstone with a characteristic conchoidal fracture. Sandstone bands locally present, especially near contact with the overlying Aotea Sandstone.	46	
			Glen Massey Formation		
			Glen Massey Sandstone Member : Interbedded calcareous silty fine sandstones and sandy siltstones.	37	
			Dunphail Siltstone Member : Blue-grey calcareous siltstone, glauconitic at base.	10	
			Elgood Limestone Member : Crystalline slightly glauconitic flaggy limestone with a highly glauconitic sandstone.	8	
	Lwh		Mangakotuku Siltstone : Massive non-calcareous siltstone, often more sandy near base, weathers to a dark ochre brown. Locally concretionary.	65	
Upper Eocene	Ar		Waikato Coal Measures : Coal measure interbedded with fireclays, brown-grey mudstones or siltstones, rarer sandstones and conglomerates. Locally, upper beds are sideritic.	148	
Mesozoic	Jurassic		Siltstones, sandstones and conglomerates with tuffaceous beds and rare concretions. Less indurated than the Triassic rocks.	6100	Average thickness
	Triassic		Indurated sandstones and siltstones, conglomerates near base, with tuffaceous beds.	2900	

Table 3.2. Stratigraphic sequence of exposed Mesozoic and Tertiary sediments in the Hamilton District (after Kear, 1960).

		N.Z. Stages			Thickness (in metres)	
Pliocene		Wo–Wm		Tangahoe Mudstone: Massive blue-grey sandy mudstone with scattered concretions, thin sandstone bands in topmost beds.	610	Thickness in Type Section area
		Tk–Wo		Matemateaonga Sandstone: Sandstone and some mudstone with numerous conglomeratic shell beds, and some concretionary bands.	270	
Miocene	Upper	Tt	Mohakatino Group	Urenui Siltstone: Massive blue-grey micaceous siltstone with some concretionary layers and rare thin tuffaceous sandstone bands. Some conglomeratic bands containing pebbles from a westerly source, and slumped beds. Eastwards away from Type area, characteristic features are lost and siltstones become indistinguishable from Mt. Messenger and Mohakatino sediments.	460	
		Tt		Mt. Messenger Sandstone: Banded mudstone with occasional tuffaceous sandstone overlain by argillaceous sandstone with concretionary bands. Eastward the sediments grade into a sandy claystone that is difficult to distinguish from Urenui and Mohakatino sediments.	730	
	Middle	Tt		Tawariki Claystone: Massive blue-grey mudstone with sandstone and thin tuffaceous bands that decrease in abundance to the east.	60	
				Ferry Sandstone: Greenish non-tuffaceous micaceous sandstone with some claystone bands.	60	
				Omoao Formation: Massive tuffaceous micaceous sandstones and interbedded blue-grey micaceous sandstones and mudstones containing variable amounts of volcanogenic materials and some concretions. Defined by Happy (1971) to describe Ferry and Tawariki sediments in the Awakino area.	76	
				Purupuru Tuffaceous Sandstone: Tuffaceous mudstone (15 m) locally containing tuffaceous sandstone bands, overlain by tuffaceous mudstone (30 m). Locally contains lenses of tuffaceous foraminiferal limestone up to 9 m thick.	46	
		Sc		Rangarara Sandstone: Grey sandstone with thin conglomerate at base, locally containing limestone lenses up to 30 m thick.	10	
	Lower	Pl	Mokau Group	Upper Mokau Sandstone: Well bedded, blue-grey, argillaceous sandstone with conglomeratic and shelly lenses and a few concretions.	170	
				Maryville Coal Measures: Upper and lower coal seams and carbonaceous shale zones separated by 15 m of massive carbonaceous sandstone.	60	
				Lower Mokau Sandstone: Massive argillaceous sandstone with some mudstone laminations and concretions. Lower part carbonaceous or with thin coal seams.	120	
		Pl Lw	Mahoenui Group	West Taranaki Taumatamaire Formation Massive blue-grey spheroidally weathered calcareous mudstone with some concretionary and sandstone bands. At western edge of depositional basin mudstone give way in part to argillaceous calcareous sandstone and limestone layers (the Upper and Lower Awakino Limestone). Sediments up to 210 m thick in Awakino Valley. East Taranaki Taumarunui Formation Alternating sandstone/mudstone flysch-type beds up to 610 m thick.		
Oligocene		Lw	Te Kuiti Group	Otorohanga Limestone: Pure white crystalline flaggy limestone containing pebbly layers near top.	6	Thickness in Awakino Area
				Orahiri Limestone: White crystalline limestone with oyster beds that sometimes show sandy, flaggy, or greywacke pebble horizons.	30	
				Mangapehi Sandstone: Hard blue-grey calcareous sparsely fossiliferous fine sand. Occasionally glauconitic or concretionary.	76	
				Aotea Sandstone: Grey calcareous fine – medium sandstone. Locally contains the flaggy Waitetuna Limestone member.	60	
		Lwh		Whaingaroa Siltstone: Massive blue-grey calcareous siltstone with a characteristic conchoidal fracture. In Awakino area unit contains the Awamarino Limestone member near base.	185	
Upper Eocene		Ar		Waikato Coal Measures: Interbedded coal measures, carbonaceous claystones, sandstones and conglomerates. Rare sideritic concretions.	15	
Mesozoic		Jurassic		Sandstones and mudstones, with grit, conglomerate, plant remains and tuff. Less indurated than the Triassic rocks.	3,700	Average thickness
		Triassic		Sandstones and mudstones with some grit, conglomerate, plant remains and tuff.	4,300	

Table 3.3. Stratigraphic sequence of exposed Mesozoic and Tertiary sediments in the Taranaki District (after Hay, 1967; Happy, 1971).

		N.Z. Stages			
QUATERNARY	Holocene			Includes river sands and silts, dune sands which are rich in titaniferous magnetite to the west of Wanganui, river terrace sediments composed of boulders, coarse gravel and sand, and unsorted andesitic material of lahar flows. Mt. Egmont derived volcanic ash mantles much of the area.	Dominantly non-marine
	Pleistocene	Hawera	Up.	St. Johns and Papaiti Formations : Dominantly non-marine gravels sands and silts.	
				Rapanui Formation : Basal conglomerate overlain by marine sands, dune sands, lignite bands, with laharic debris.	Partly marine
			Lr.	Brunswick Formation : Andesitic gravels, marine sands, pumice, clay and dune sands. Kaiatea Formation : Quartzite, argillite and greywacke gravels, volcanic ash and dune sand.	
		Wc		Shell conglomerate, siltstones and shell grit overlain by pumiceous sand and siltstone with shell beds and shell grit. Topmost beds composed of alternating sands shell beds and siltstones.	Dominantly marine
		Wn		Upper Wn : Blue-grey sand, silt and clay with non-marine beds, capped by non-marine current bedded sands, silts, clays and lignite. Lower Wn : Fine to medium grained, brown micaceous sands with numerous coquina limestone bands and some concretionary muddy sands.	
TERTIARY	Pliocene	Wm		Massive mudstone, shell conglomerate and muddy sands grading into loose sands with lenticular shell beds, overlain by micaceous sands and numerous shell beds.	Marine
		Wp		Complex sequence of interbedded mudstone, coarse to fine sands and numerous shell beds.	
		Wo		Blue-grey massive mudstones overlain by massive medium to fine cross-bedded sands with lenticular shell beds.	

Table 3.4. Stratigraphic sequence of exposed Tertiary and Quaternary sediments in the Wanganui District (after Lensen *et al.*, 1959).

3.1.2 PHYSIOGRAPHY

The major physiographic features of the onshore study area are summarised in Fig. 3.2.

The Hamilton District is bordered to the west by high (300-600 m) coastal ranges formed in well dissected Mesozoic rocks and Quaternary volcanic cones (Fig. 3.2). Tertiary rocks, moderately to well dissected, lap on to the coastal ranges at lower levels. To the southeast the eroded Mesozoic rocks of the Rangitoto Range rise to 900 m, and are flanked at lower levels by ignimbrite sheets and further to the west by Tertiary sandstones and limestones. The northeastern part of the Hamilton Lowland is bordered to the west, north and east by ranges cut in Mesozoic sandstones and mudstones and contains low lying Quaternary rhyolitic sands and gravels with some peat deposits.

The central depression of the Taranaki District is bordered to the northwest by the rugged Herangi Range (800 m) and to the northeast and east by the Hauhungaroa Range (750-1100 m), both ranges being cut in Mesozoic rocks. The lower well dissected plateau between these ranges dips gently to the south and west, forming the northern part of the Taranaki Basin. In the north and east the interfluves of Tertiary rocks are capped by remnants of a once extensive ignimbrite sheet. The southern topography is characterised by deeply entrenched rivers. In the southeast a lahar plain reaches out from the Central Volcanic Region and to the southwest a chain of andesitic volcanoes rise above a plain dotted with low hills formed by erosion of volcanic mud flow debris.

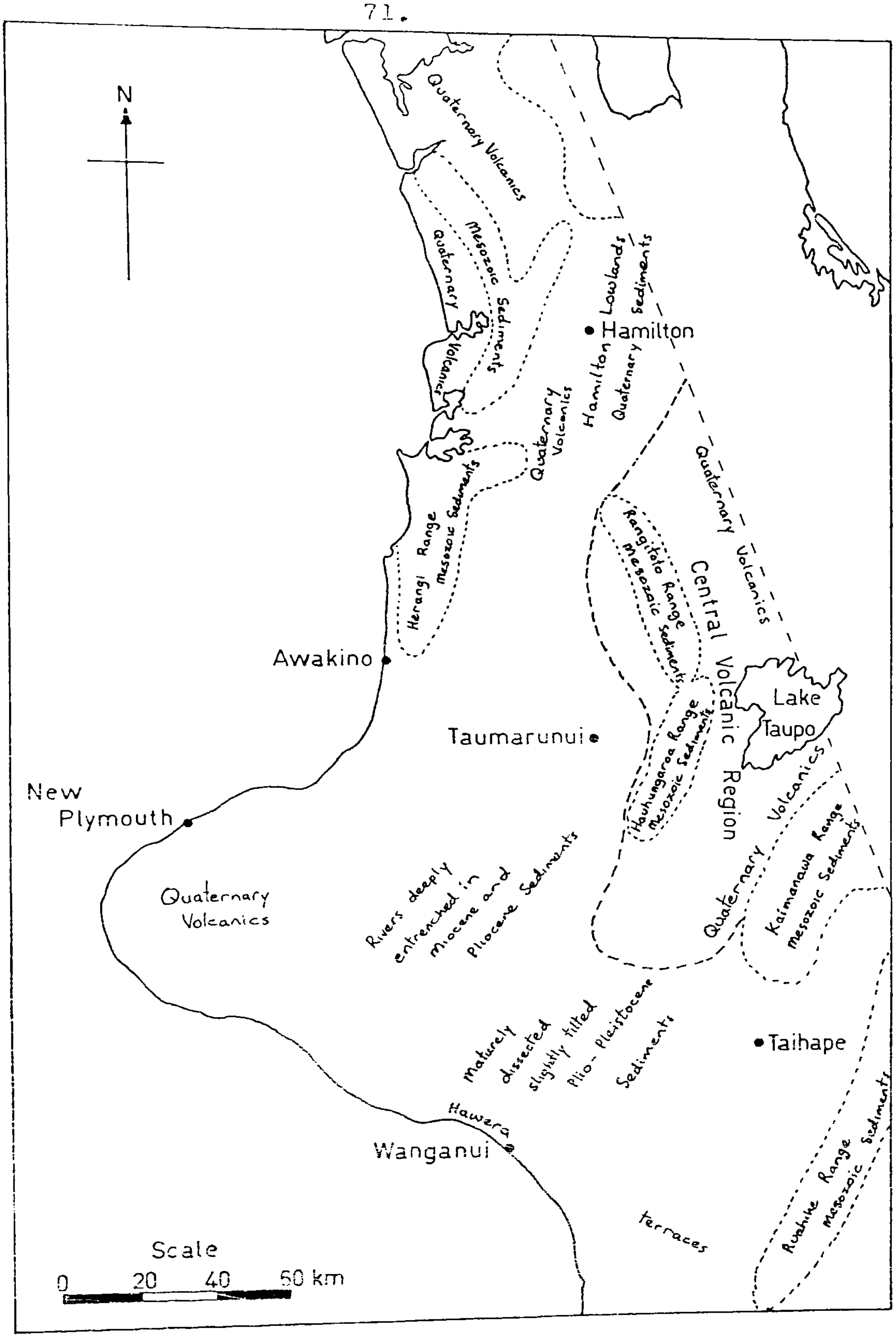


Fig. 3.2. Major physiographic features of the onshore study area (after Lensen *et al.*, 1959; Kear, 1960; Hay, 1967).

The uplands of the Wanganui District consist of well dissected cut surfaces with an alluvial veneer and maturely dissected, slightly tilted, Plio-Pleistocene marine sediments. South of Wanganui the Upper Pleistocene (or Hawera) aggradational surfaces slope gently seawards and form the coastal lowlands.

3.1.3 STRUCTURE

The major structural elements of the onshore study area are summarised in Fig. 3.3.

The western Mesozoic rocks of the Hamilton District are broadly folded into the regional north-trending Kawhia Syncline, a structure cut by faults running parallel to the synclinal axis and bordered to the east by the major Waipa fault. The eastern ranges of Mesozoic rocks are part of a complex anticlinal structure that is fault-bounded on its western border. Tertiary and Quaternary sediments of the Hamilton Lowlands are either flat-lying or very broadly folded and block-faulting is common (Kear, 1960).

In the Taranaki District the Herangi Range, a structural extension of the Kawhia Syncline to the north, probably extends southwards under Tertiary rocks. The structure of the eastern ranges is uncertain. The stratigraphy of the sediments in the northern Taranaki District differs from that in the southern Hamilton District (cf. Tables 3.2 to 3.4), the two regions being separated by a shallow threshold near Piopio named the Piopio Threshold (Hay, 1967). Structurally the Taranaki District has been described as "... a much faulted monocline dipping south and west and divided into two parts by the north-south

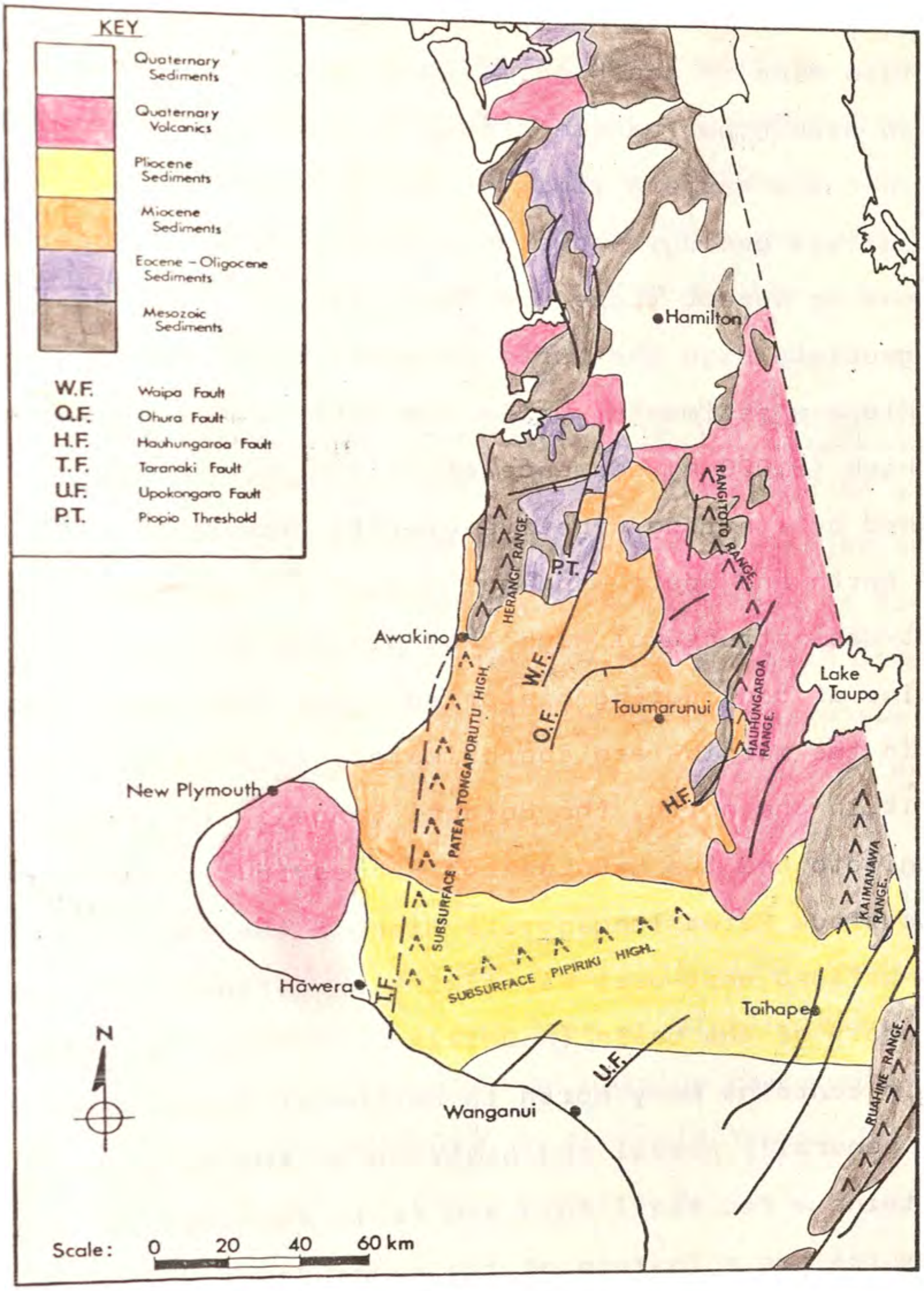


Fig. 3.3. Major structural features and generalised geology of the onshore study area (after Lensen et al., 1959; Kear, 1960; Hay, 1967).

trending Patea-Tongaporutu Gravity High" (Hay, 1967). The western side of the High is bounded by the Taranaki Fault, which downthrows basement 6000 to 7500 m to the west and defines the eastern limit of the Taranaki Basin. Here, subsurface geology shows an almost complete succession from Eocene to Recent (Cope and Reed, 1967). East of the Patea-Tongaporutu High the North Wanganui Basin contains Eocene to Miocene sediments, while the South Wanganui Basin shows a thick (5,000 m) sequence of Plio-Pleistocene sediments marked by a strong negative gravity anomaly (Reilly, 1965). The North and South Wanganui Basins are separated by an east-west structural high that divides discrete sedimentary basins of contrasting histories. For the North Wanganui Basin the margins are approximately defined by the Maracopa Fault in the north, the Herangi Range in the northwest, the Rangitoto and the Hauhungaroa Ranges in the east, the subsurface Patea Tongaporutu High in the southwest and the ill-defined east-west Pipiriki High defines the southwestern boundary of the basin (McQuillan, 1977). The North Wanganui basin contains many north to northeast trending faults which are generally normal and upthrown on the northwest, and contains a few small fold and fault structures. Local flexures are a feature of the major fault zones, those along the Ohura Fault being especially prominent. The Patea-Tongaporutu High is interpreted as a horst-like feature which emerged in the Oligocene and was active in the Miocene and Pliocene. During Mohakatino, Mt. Messenger and Urenui times, the Patea-Tongaporutu High influenced sedimentation, and different sediment types accumulated on the flanks (Hay, 1967; Nelson and Hume, 1977).

The sediments of the Wanganui District, occupying the South Wanganui Basin, have a simple regional structure, dipping southwest towards Cook Strait and thickening eastwards. The Wouth Wanganui Basin is a complex graben consisting of secondary horsts and grabens over which the Tertiary and Quaternary strata drape in gentle anticlines and synclines (Lensen et al., 1959).

3.2 COMPOSITIONAL CHARACTERISTICS OF THE SEDIMENTS

Sediment compositional data is presented in a manner that shows the petrographic character of the major sedimentary units in the onshore study area. Figures 3.4A and 3.4B, and 3.5 and 3.6 summarise the average bulk mineral and clay mineral composition of the sediments, while deviations from these average values are detailed in Appendix Tables III.1 and IV.1 and described in the text.

3.2.1 BULK MODAL COMPOSITION

The distribution of quartz, plagioclase, potash feldspar, and clay minerals in the acid insoluble residues of sediments and the distribution of calcium carbonate in bulk sediment samples are detailed in Appendix Tables III.1 and III.2 and summarised in Figs. 3.4A and B.

3.2.1a QUARTZ

Quartz occurs in only small quantities in all limestone lithologies. In mudstones quartz averages 10-30% but increases in abundance in basal Te Kuiti (i.e. Waikato Coal Measures and Mangakotuku Siltstone), lower Mokau Group and Maryville Coal Measures sediments. The absolute abundance

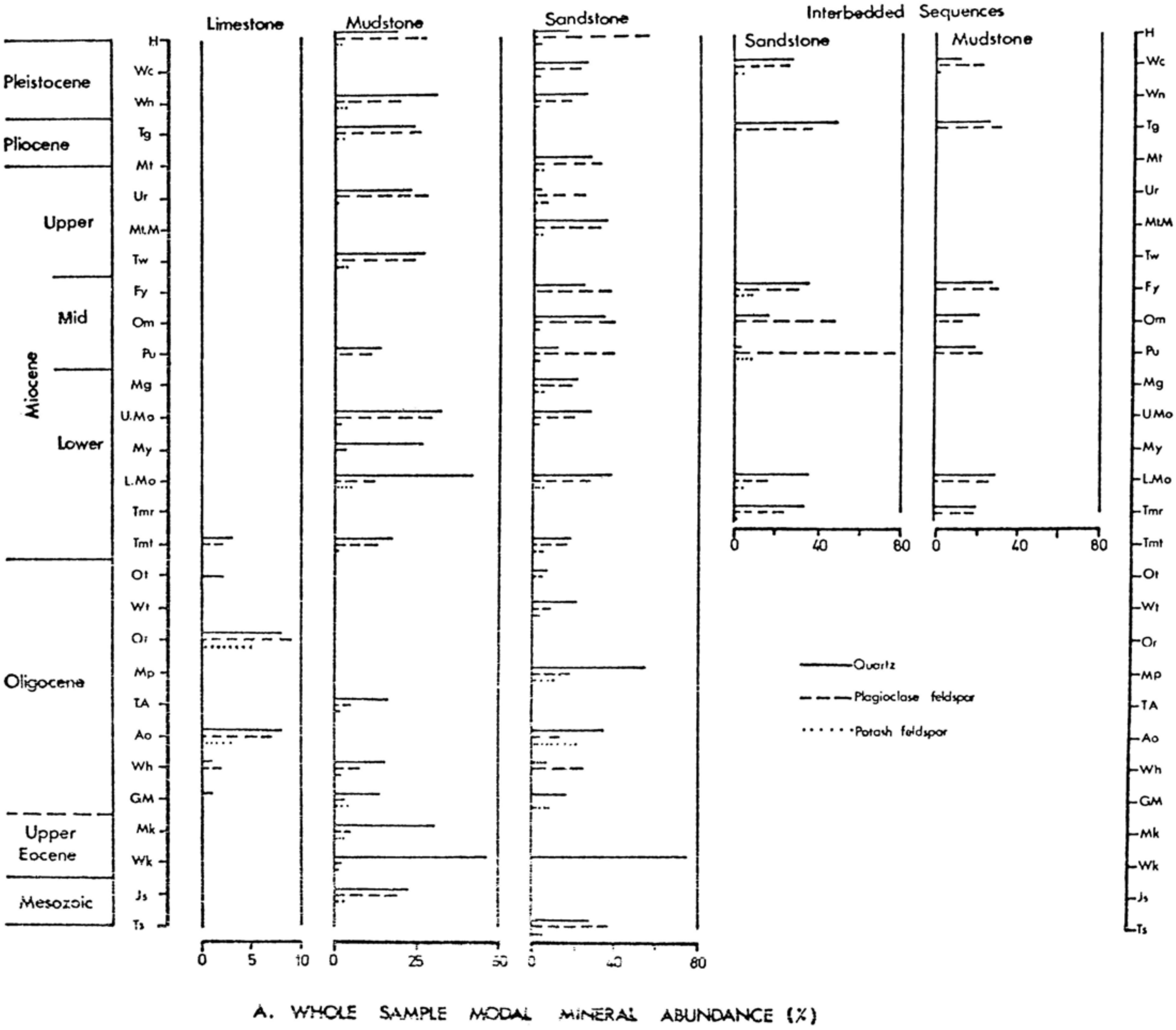


Fig. 3.4. Bulk variations in average mineral compositions for the major sedimentary units in the onshore area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1. Additional "lift-out" copy available in back pocket.

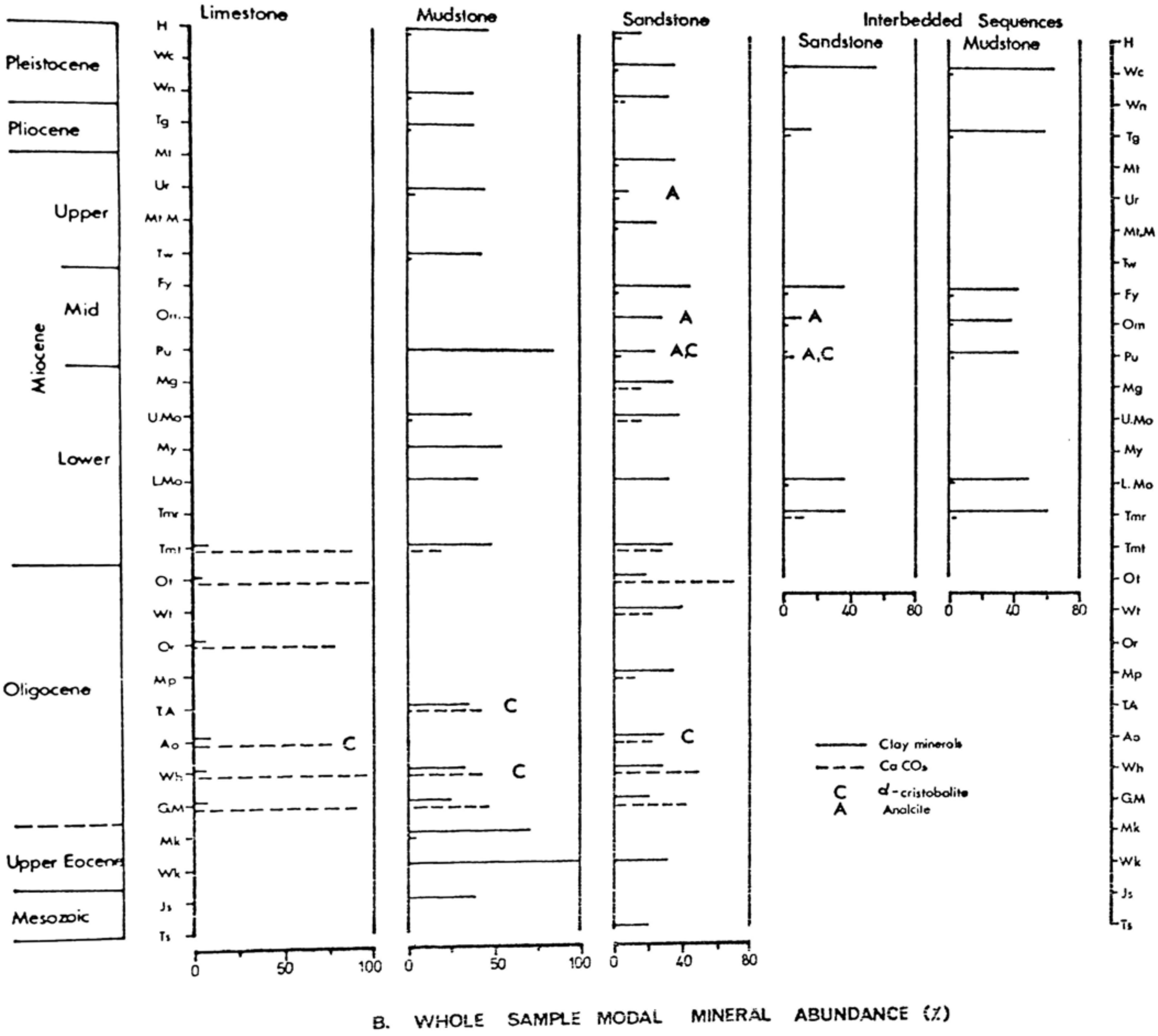


Fig. 3.4. Bulk variations in average mineral compositions for the major sedimentary units in the onshore study area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1. Additional "lift-out" copy available in back pocket.

of quartz is generally low in the Te Kuiti Group because of the highly calcareous nature of the sediments; quartz, however is a major component in the insoluble residues of the Te Kuiti Group sediments. The quartz distribution in sandstones is more complex, averages 10-30%, but increases locally in massive sandstones (e.g. L.Mo: 10275, 10276, 10277, 10281; Om: 10303; Mt. M: 10317, 10318, 10319), leached sandstones (e.g. Ao: 10226) and on-basement samples (e.g. Wk: 10205, 10238, 10239). The quartz distribution in interbedded sandstone/mudstone sequences parallel quartz distribution in the massive sandstone and mudstone units. Within the major lithologic type in any one formation the quartz abundance varies by about ± 5% (Appendix Table III.1), but shows local increases in on-basement lithologies and in leached sandstones.

3.2.1B PLAGIOCLASE FELDSPAR

The distribution of plagioclase roughly parallels that of quartz with a few notable exceptions. The relative abundance of plagioclase to quartz decreases markedly in lower Te Kuiti and lower Mokau Group sandstones and mudstones but is generally subequal in abundance to quartz throughout Te Kuiti sedimentation. The plagioclase abundance increases in volcanogenic sandstones (i.e. Purupuru, Omoao, Urenui and Hawera) and in Recent ashes, where X-ray diffractograms indicate the presence of two types of plagioclase feldspar (A and B) of subequal abundance. The variations in plagioclase abundance within individual lithologic units of a formation are less than for quartz (i.e. < ± 5%).

3.2.1C POTASH FELDSPAR

Potash feldspar is of low abundance in most samples. It forms over 15% of the acid insoluble residue in a number of samples from a location (i.e. 28) adjacent to the Herangi Range (e.g. Or: 10240, 10241; Ao: 10225, 10226, 10230).

3.2.1D CLAY MINERALS

In limestones clay minerals are more abundant than either quartz or feldspar. Massive mudstones contain from 35-45% clay minerals, but this value increases in volcanogenic mudstones (e.g. Purupuru, and Omoao), in the Mangakotuku siltstones, in the Waikato Coal Measure mudstones, in the Maryville Coal Measures mudstones and in the Mahoenui sediments (Appendix Table IV.1). Clay minerals constitute 30-40% of most massive sandstones but volcanogenic sandstones (e.g. Purupuru, Omoao, Urenui and Hawera) and the Triassic sandstones are notably deficient in clay minerals. Interbedded sandstones and mudstones show generally similar distributions in clay mineral abundances as their more massive time-correlatives but, compared to sandstone lithologies, clay mineral abundances in the mudstones are notably greater. Clay mineral abundance can vary markedly between samples from any one formation.

3.2.1E CALCIUM CARBONATE

With the exception of the Waikato Coal Measures, the Mangakotuku siltstones, and certain leached sandstones (e.g. Ao: 10225), the Te Kuiti Group sediments are characterised by high $CaCO_3$ contents. Smaller quantities of $CaCO_3$ occur in Mahoenui, Upper Mokau, Mangarara and Nukumaruan sediments. Concretions contain from 30-80% $CaCO_3$, with highest values in the older concretions.

3.2.1F OTHER MINERALS

Low temperature cristobalite occurs in some samples of Whaingaroa Siltstone, Aotea Sandstone and Te Akatea Siltstone (Appendix Table IV.1 and Fig. 3.4B), forming up to 10% of the sample. In all the older volcanogenic sequences (i.e. Purupuru, Omoao and Urenui) both analcite and α-cristobalite are characteristic. Analcite constitutes some 40% of samples in the Urenui volcanogenic sandstones.

3.2.2 CLAY FRACTION COMPOSITION

The distribution and crystallinity of clay minerals in the <2μ fraction of sediments in the study area are detailed in Appendix Tables IV.1, IV.2 and IV.3 and summarised in Figs. 3.5 and 3.6.

3.2.2A KAOLINITE

Kaolinite occurs in only small quantities in most samples. It is absent from Triassic sediments and rare in Jurassic samples. In the basal Te Kuiti Group carbonaceous mudstones and sandstones, kaolinite is the dominant clay mineral species (Pl. 3.4) and persists in abundance into the overlying Mangakotuku siltstones (Pl. 3.6). The younger Te Kuiti mudstones, sandstones and limestones contain no or rare kaolinite, although local enrichments occur in on-basement lithologies (e.g. Ot: 10245; Mp: 10238, 10239), in samples adjacent to Mesozoic basement highs (e.g. locality 28), and in leached sandstones (e.g. Ao: 10226).

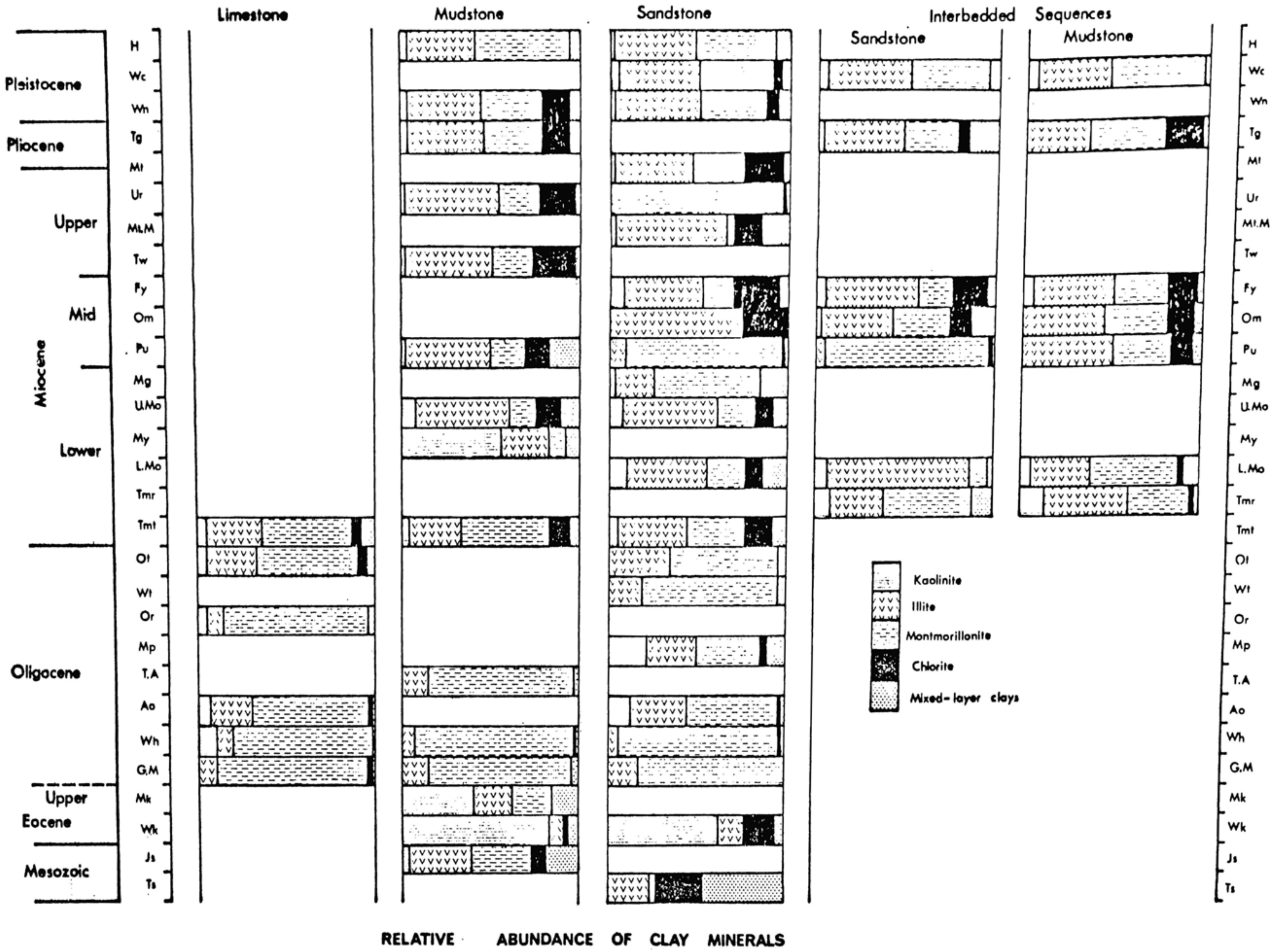


Fig. 3.5. Bulk variations in average clay mineral compositions for the major sedimentary units in the onshore study area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1. Additional "lift-out" copy available in back pocket.

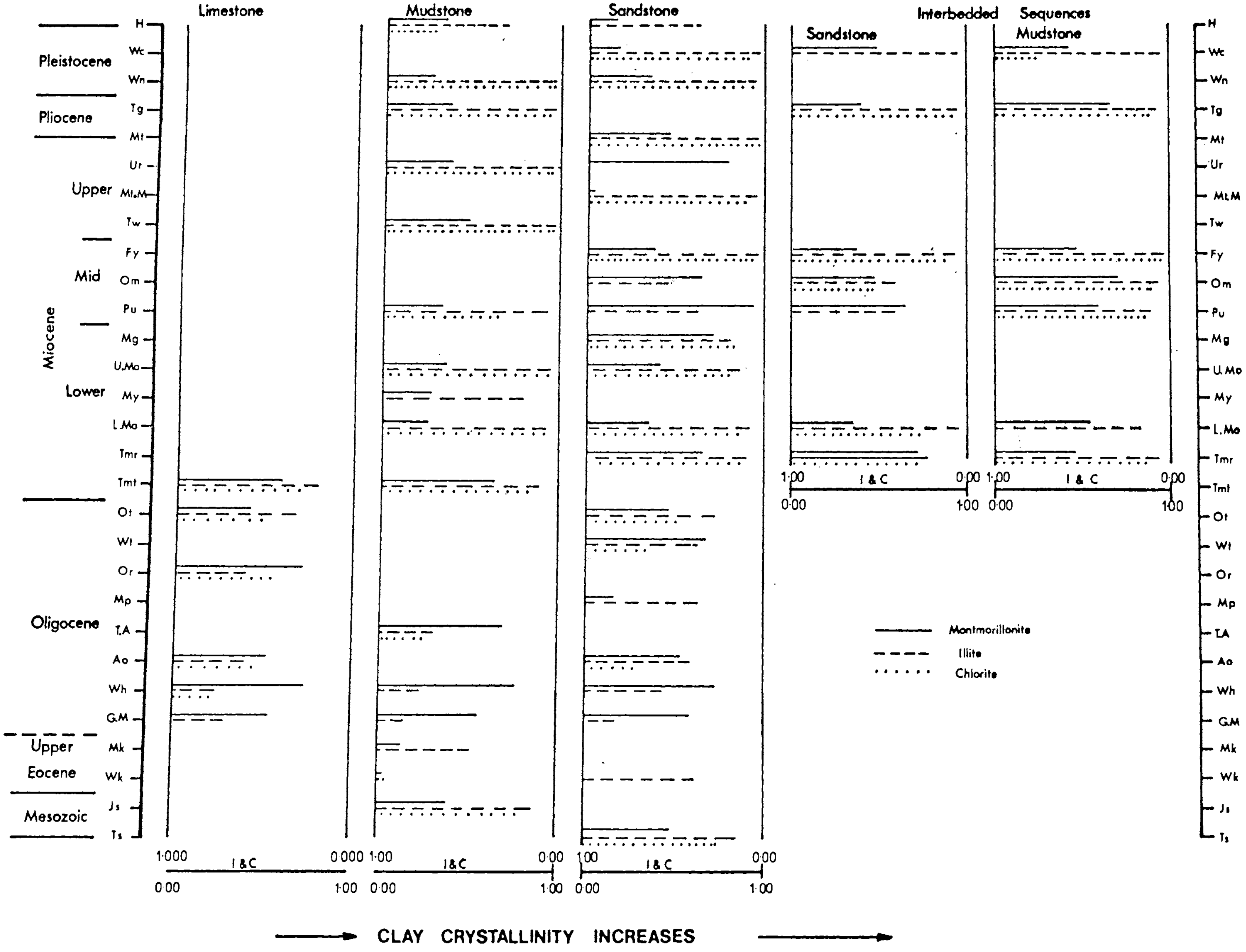


Fig. 3.6. Bulk variations in average clay mineral crystallinity for the major sedimentary units in the onshore study area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1. Additional "lift-out" copy available in back pocket.

Kaolinite constitutes less than 10% of most Mahoenui samples. The Taumatamaire Formation has less than 5% kaolinite, except in on-basement lithologies where values reach 10% (e.g. 10252, 10258, 10260). In the flysch beds of the Taumarunui Formation however, the kaolinite content may be as high as 10-20%.

Lower and Upper Mokau sandstones and mudstones are kaolinite poor (<5%) except for samples from on-basement lithologies (e.g. L.Mo: 10278), from the carbonaceous Maryville Coal Measures (e.g. My: 10282) from close to major fault systems (e.g. Aria Fault L.Mo: 10278; Ohura Fault L.Mo: 10275), and from shore-line facies (e.g. L.Mo: 10281; U.Mo: 10290, 10291).

Throughout Mangarara, Purupuru and Omoao times, kaolinite is virtually absent. In upper Mohakatino times it constitutes less than 5% of the clay fraction, increasing locally to 10%. The Upper Miocene to Pleistocene sediments investigated in this study, contain little kaolinite (<5%).

The X-ray diffractograms indicate that the kaolinites from the coal measures are more crystalline than those from leached sandstones and soils, a property supported by heating characteristics.

3.2.23 ILLITE

Illite is an important component of the clay fraction in most of the sediments studied (Fig. 3.5). In Triassic and Jurassic sediments it forms some 25% of the clay fraction. The lower Te Kuiti mudstones contain only small amounts of illite (<10%) but in the upper Te Kuiti lithologies the illite content increases to 20-25%. Illite is locally more abundant in the Te Kuiti sediments including for example, in some Mangakotuku siltstones, (e.g. 10214), in certain less intensely leached sediments of the Waikato Coal Measures (e.g. 10207, 10208), in mudstones and limestones from on-basement locations (e.g. Ao: 10223; Ot: 10245, 10246), in intensely leached sandstones (e.g. Ao: 10226), and in glauconitic sandstones (e.g. Ao: 10225, Ot: 10244).

Above the Te Kuiti Group the amount of illite in the clay fraction of sediments increases markedly. The Taumatamaire mudstones contain 20-25%, increasing to 30-50% adjacent to basement highs such as the Herangi High. To the east the Taumarunui flysch beds contain from 20-60% illite with highest values in the northern area. In the lower Mokau beds the sandstones contain 30-50% illite, whereas the interbedded sandstones and mudstones contain 80% and 35% illite respectively. In the Upper Mokau mudstones the illite content reaches 50-60%, but above this level and through to the Recent the illite abundance in mudstones decreases to 40%.

In general terms illite forms 40-50% of the clay fraction of most sandstones, but increases to 60-70% in massive sandstones (e.g. Om: 10303; and Mt. M: 10317, 10318, 10319) and decreases to 0-5% in volcanogenic sandstones (e.g. Purupuru and Urenui). The illite content in interbedded sandstone/mudstone sequences shows similar trends to their more massive lithologic counterparts. As a general rule sandstones contain more illite than mudstones.

High crystallinity characterises the illite in the oldest (Mesozoic) sediments (Fig. 3.6). Most Te Kuiti Group sediments contain illite of low crystallinity, although the Mangakotuku siltstones and the youngest Te Kuiti sediments contain more crystalline illite. From Mahoenui times onwards the illite crystallinity in mudstones is high, but by the end of the Pleistocene the illites are poorly crystalline. The illites in sandstones exhibit generally similar trends except that in volcanogenic sequences their crystallinity is very low. The illites of interbedded sandstone/mudstone sequences show similar trends to their more massive age correlatives.

Glauconitic samples commonly show an abundance of illite (e.g. Ao: 10225). This is due to the coincidence of the major XRD peaks of illite and glauconite, the latter mineral being essentially an iron-rich illite (Burst, 1958). Samples containing glauconite in the clay fraction will therefore show a corresponding increase in illite.

3.2.2C MONTMORILLONITE

Montmorillonite is rare in Triassic sandstones (< 5%) but is an important component of the clay fraction of some Jurassic mudstones (e.g. 10204) and concretions (e.g. 10202).

The Waikato Coal Measure sediments contain only minor montmorillonite but montmorillonite increases in abundance in the Mangakotuku siltstones (0-30%). In the remainder of the Te Kuiti sediments and extending through into the Mahoenui Group montmorillonite is the dominant clay mineral species. In limestones and mudstones, and to a lesser extent sandstones, montmorillonite constitutes 50-90% of the clay fraction, decreasing only in on-basement samples (e.g. Ao: 10233; Ot: 10245, 10246) which show a concomitant increase in illite abundance.

The Taumatamaire mudstones are rich in montmorillonite, especially in the north (70% montmorillonite), but to the west, near the Herangi High, the montmorillonite content drops to 20-40%; the limestones adjacent to the high, however, retain 40-70% montmorillonite. The graded Taumarunui sediments contain variable amounts of montmorillonite but, in general, sandstones contain more montmorillonite in the clay fraction than do mudstones, and the montmorillonite content is highest in the south.

The abundance of montmorillonite continues to decrease through the Lower Mokau sediments (20% montmorillonite), but samples from the northern extremities of the formation retain a higher montmorillonite content (35-55%). The Upper Mokau mudstones and sandstones contain approximately 20% montmorillonite in their clay fraction except for one sample (i.e. 10286) collected near the Patea-Tongaporutu High in which the montmorillonite content is low (< 5%). In general mudstones from nearshore facies are poor in montmorillonite (e.g. 10290, 10291).

From Mohakatino times until the Recent the massive mudstones, their enclosed concretions, and the mudstones from flysch-type sequences show a gradual increase in montmorillonite. All volcanogenic sandstones (e.g. Mangarara, Purupuru, Urenui and Hawera) have high montmorillonite contents (70-95%). Mangarara sandstones show higher montmorillonite contents in the west (80%) than in the east (40%).

Non-volcanogenic upper Mohakatino mudstones and sandstones contain 20-30% montmorillonite. The abundance of montmorillonite declines markedly in the Mt. Messenger sandstones (5% montmorillonite), but in younger sediments montmorillonite again increases in abundance (e.g. Urenui mudstones 20-30%, Matemateaonga sandstones 30%, Tangahoe mudstones 30-50%, and Hawera sandstones 40-50%). The Plio-Pleistocene increase in montmorillonite content is reflected in all lithologies.

Both Jurassic and Triassic sediments contain montmorillonite of moderate crystallinity (Fig. 3.6). The crystallinity of montmorillonite in the basal Te Kuiti sediments (i.e. Waikato Coal Measures and Mangakotuku siltstones) is low, but for the remainder and bulk of the Te Kuiti lithologies and in the Mahoenui Group, the crystallinity is moderate to high. In the Mokau sediments montmorillonite crystallinity suddenly decreases and, in the mudstones at least, remains constantly low until Recent times. The crystallinity in sandstones shows a generally similar trend but is complicated by a sharp increase in montmorillonite crystallinity in volcanogenic sandstones (e.g. Mangarara, Purupuru, Omoao, and Urenui), and a pronounced decrease in montmorillonite crystallinity in massive, porous sandstones (e.g. Omoao and Mt. Messenger) and in unconsolidated, porous sands (e.g. Castlecliffian and Hawera).

X-ray diffractograms indicate that most of the montmorillonites investigated contain Ca^{2+} and/or Mg^{2+} as the exchangeable cation associated with two layers of water, and exhibit disordered, irregular interstratification. Some basal Te Kuiti montmorillonites (e.g. Mk: 10214, G.M : 10216, 10217, 10218) may contain Na^+ as the exchangeable cation associated with one layer of water. In general, the montmorillonite of volcanogenic sandstones is trioctahedral, probably stevensite, while the montmorillonite in all other sediments is a dioctahedral species, probably montmorillonite (Appendix Table VIII.2). Treatment of selected samples with KOH (Table 2.6) indicates that the montmorillonites were probably not derived from the alteration of micaceous materials (see p. 43).

3.2.2D CHLORITE

Chlorite is a major component (30%) of the clay fraction of Triassic sediments but is less abundant (10%) in the Jurassic mudstones studied. In the Te Kuiti sediments chlorite is present in only trace amounts (< 5%) but increases locally in on-basement lithologies (e.g. Mp: 10239), in some of the less intensely leached fireclays of the Waikato Coal Measures (e.g. 10205), and in some Otorohanga limestones (e.g. 10247, 10248).

In Mahoenui Group and younger sediments the amount of chlorite in the clay fraction tends to increase. Taumatamaire mudstones contain less than 10% chlorite but mudstones and sandstones from locations adjacent to the Herangi Range contain more chlorite (10-30%). Awakino limestones contain little (5-10%) chlorite.

The abundance of chlorite in Lower Mokau sediments varies from 0-30%, but the larger amounts, as was the case for kaolinite, are confined to those sediments adjacent to active fault zones (e.g. 10278, Aria Fault; 10275, Ohura

Fault). Chlorite persists in amounts of 10-20% in most sediments until middle-Pleistocene times, after which it becomes rare. Exceptions to this general rule are as follows: (1) no, or only trace amounts of chlorite occur in volcanogenic sandstones, and interbedded volcanogenic mudstones contain only 10-15% chlorite; (2) chlorite increases in abundance (25-30%) in sediments from areas of uplift (e.g. OM: 10299, 10301; adjacent to Herangi Range), and in sediments adjacent to major fault systems (e.g. Fy: 10305, 10309, 10312 and Tu: 10316; adjacent to Ohura Fault).

The crystallinity curves for chlorite parallel those of illite, although the chlorite is usually less crystalline than illite. Chlorites from the Mesozoic and Te Kuiti sediments are the heat-labile varieties (Fig. 3.7). Stable chlorites first appear in the Taumatamaire mudstones and sandstones adjacent to the Herangi Range, but do not occur in the other Mahoenui sediments. The stable species of chlorites, along with heat-labile chlorite, persist in sediments until the end of the Nukumaruan. From Castlecliffian times onwards chlorites are of the heat-labile and mixed-layer varieties. XRD analyses, indicate that the chlorite in Triassic samples is a Mg-rich variety, while in all other sediments the chlorites have a large Fe/Mg ratio.

3.2.2E MIXED-LAYER CLAYS

The distribution of mixed-layer clay mineral species in the clay fraction of sediments is presented in Appendix Table IV.1 and summarised in Figs. 3.5 and 3.7. Mixed-layer clay minerals (mostly C-V) are a major component (45%) of the clay fraction of Triassic sediments but are less abundant

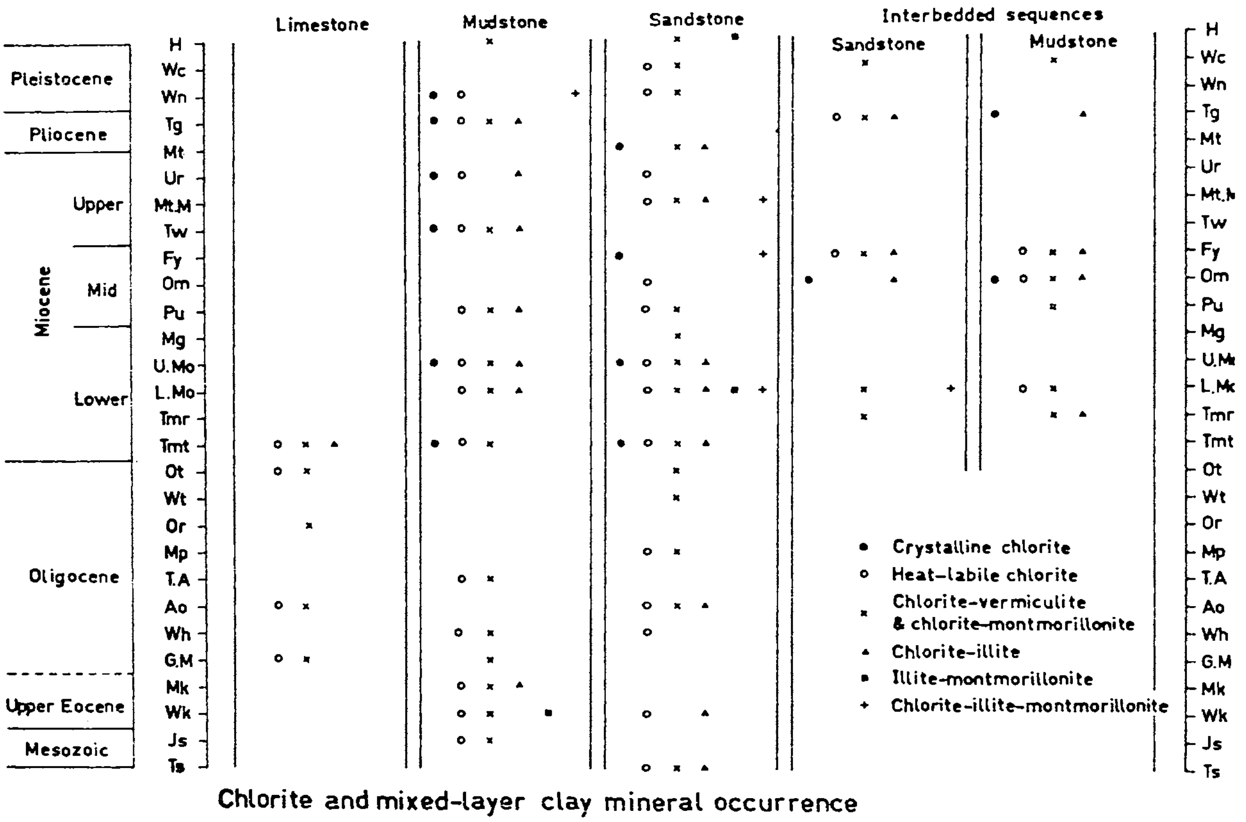


Fig. 3.7. The occurrence of chlorite and mixed-layer clay mineral species in sediments of the onshore study area. Abbreviations of formation names defined in Table 1.1.

(20%) in Jurassic samples. In the Te Kuiti and Mahoenui Group sediments mixed-layer clays (mostly C-V and C-M) are of minor importance (0-10%) but increase in abundance in sediments of nearshore facies (e.g. Mangakotuku siltstone), in Taumatamaire sediments and Mokau sediments from locations adjacent to the Herangi Range (10-15% mixed-layer clays), and in the Middle and Upper Miocene formations (approximately 20%). Volcanogenic sandstones, however, contain only trace amounts of these minerals. In Pliocene and Pleistocene sediments mixed-layer clay minerals are of minor importance.

3.2.2F OTHER MINERALS

Quartz is the dominant accessory mineral in the < 2μ fraction in the lower to middle Te Kuiti Group sediments. Plagioclase feldspar occurs in most samples, and in the volcanogenic sandstones (e.g. Purupuru, Urenui and Hawera) there are two plagioclase species (A + B) occurring in subequal quantities. Amphibole occurs in Te Kuiti samples and in sediments of volcanic origin. α-cristobalite is present in some Te Kuiti sediments (Whaingaroa Siltstone, Aotea Sandstone, and Te Akatea Siltstone) and in volcanogenic sediments (Purupuru Volcanic Sandstone and Omoao Formation sandstones). Analcite occurs only in volcanogenic sediments (e.g. Omoao and Urenui tuffaceous sandstones).

3.2.3 RELATIONSHIP BETWEEN SEDIMENT LITHOLOGY AND CLAY MINERAL COMPOSITION

The trends in clay mineral abundances and crystallinities shown by mudstones are paralleled by the clay minerals from limestones (Figs. 3.5 and 3.6). Sandstones, however, show quite different patterns in their clay mineral composition (Figs. 3.5 and 3.6) that may be related to differences in source materials and/or to diagenetic alteration. For example, the mudstones in interbedded mudstone/sandstone sequences generally contain more montmorillonite and chlorite relative to illite and mixed-layer clay minerals than do the sandstones. This generalisation does not apply to volcanogenic sediments. During the early deep-burial stage of diagenesis porosity is commonly high, and many types of transformation may occur as a result of circulation of formation waters

(Muller, 1967; Dunoyer de Segonzac, 1970, p.282). The association of massive porous sandstones (e.g. Mt. Messenger sandstones and Omoao sandstones) with increased abundances of mixed-layer clay minerals, together with the concomitant decrease in montmorillonite abundance, suggest that alteration of the clay minerals has occurred in these permeable sediment units. Comparison of clay mineral assemblages in the sandstones and in the mudstones of interbedded sandstone-mudstone sequences provides direct evidence of the greater amount of alteration of clay minerals in sandstones (Fig. 3.8). In particular the mudstone units are relatively enriched in montmorillonite and chlorite while the sandstones contain more kaolinite, illite and mixed-layer clay minerals. While the pore fluids in mudstones may approximate the composition

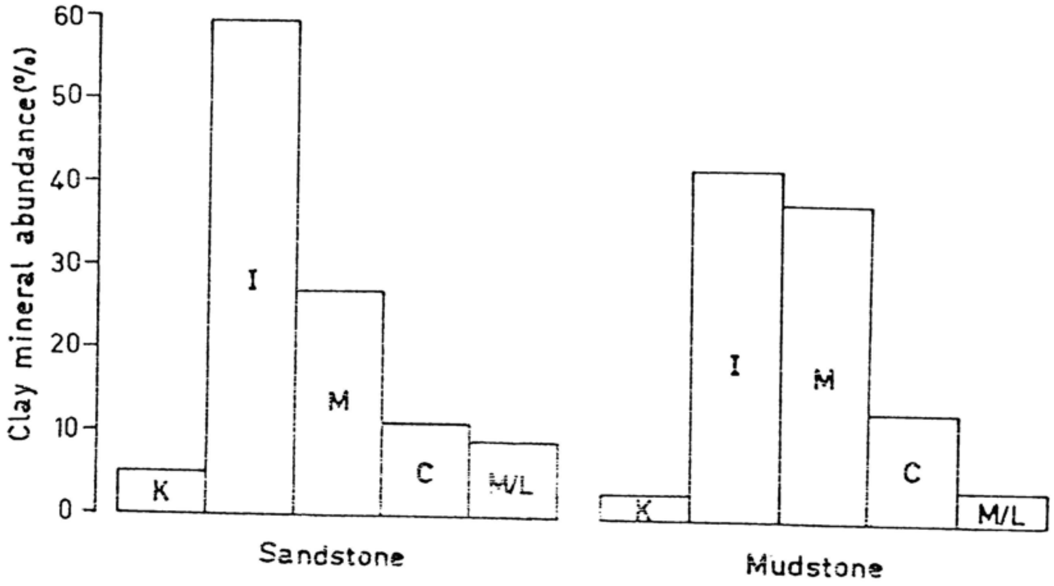


Fig. 3.8. Relative abundance of the major clay mineral species in sandstone and mudstone units of interbedded sequences (based on 18 samples). Note how the sandstone units are relatively enriched in illite, mixed-layer and kaolinite clay minerals compared to the mudstones.

of the original sediment pore waters, sandstones serve as channelways for solutions from a wide area which affect the degradation of clay material.

If the clay mineral composition of sandstone and mudstone lithologies was similar at the time of deposition, then existing differences in composition may be explained by post-depositional alteration of the clay minerals in the more porous sandstones. However until it is proven that the clay mineralogy in a sandstone sample is the same as that existing at the time of deposition of the sediment clay mineral data from sandstones must be applied cautiously in paleoenvironmental analysis. In this study a technique that involved comparing the clay mineralogy of concretions to that of their host mudstones was used with a view to determining whether or not the clay mineralogy of the mudstone was representative of that existing at the time of deposition. Concretions are most commonly formed at an early stage in the lithification of sediments (Muller, 1967; Curtis et al., 1972) and it is reasonable to assume that, once formed, their impermeable nature would protect the enclosed clay minerals from major diagenetic alteration. A comparison of the clay mineral assemblages in concretions with those from the host mudstones (Fig. 3.9) indicates that, at least since the time of concretion formation, there have been no appraciable changes in the clay mineralogy of the mudstones. For the above reasons the clay mineral assemblages in mudstones are considered to represent best those existing at the time of deposition.

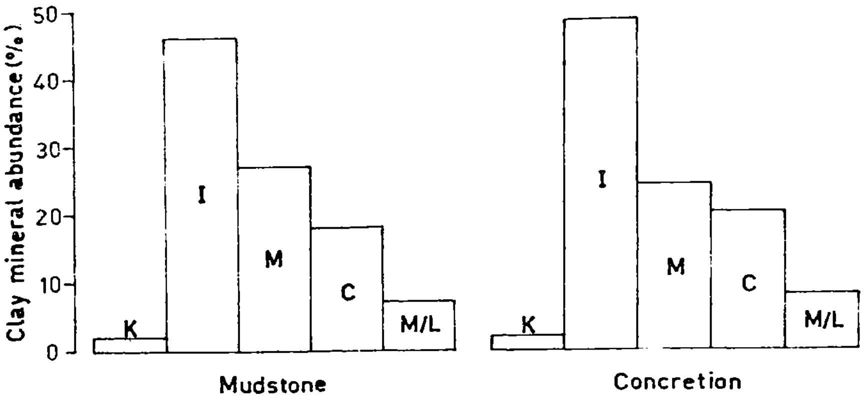


Fig. 3.9. Relative abundance of the major clay mineral species in concretions and host mudstones (based on 30 samples).

3.3 ORIGIN OF THE CLAY MINERAL SPECIES

3.3.1 MESOZOIC (TRIASSIC AND JURASSIC SEDIMENTS)

Only a small number of Mesozoic samples have been analysed and the clay mineralogy documented below for the Triassic and Jurassic sediments may not be fully representative of Mesozoic sediments generally.

The Triassic samples analysed in this study and those examined by Nelson (1973) show that illite, chlorite (magnesium-rich) and mixed-layer clay minerals (mostly chlorite-vermiculite and chlorite-montmorillonite) are the dominant clay mineral species (Fig. 3.5). Both illite and chlorite are highly crystalline (Fig. 3.6) and dominantly anhedral in form (Pl. 3.1) suggesting a detrital origin. Grain fabrics examined by scanning electron microscopy and thin section analysis show no evidence for the authigenic growth of clay minerals. The small amounts of poorly crystalline montmorillonite could be detrital from the landmass(es) that bordered the Triassic sedimentary basin(s), or they could have been derived by the diagenetic alteration of volcanogenic detritus that occurs in Triassic sediments (Coombs, 1960; Kear, 1960). Their poor crystallinity would suggest that if the latter process was operative then the montmorillonite must have suffered some degradational alteration subsequent to its formation. It is possible that the concentration of montmorillonite was originally greater than exists at present. However, the indurated nature of the Triassic sediments, the abundance of chlorite-vermiculite and chlorite-montmorillonite irregular mixed-layer clays, the high chlorite to illite ratio and the magnesium-rich nature of the chlorite (cf.p.89), all indicate that any such montmorillonite would have been transformed via chlorite-montmorillonite and chlorite-vermiculite into chlorite by diagenetic aggradation in a magnesium-rich environment. The absence of kaolinite in the Triassic sediments could be provenance controlled or result from the transformation of this mineral in a confined alkaline environment over the long period of time available (e.g. see Dunoyer de Segonzac, 1970).

In contrast to the Triassic beds, the Jurassic sediments analysed contain subequal amounts of illite, montmorillonite and mixed-layer clays and some kaolinite and chlorite (Fig. 3.5). Nelson (1973), however, suggested that montmorillonite was rare in the sediments he analysed. In as much as volcanic material is relatively common in the Jurassic sediments (Kear, 1960) it is likely that much of the montmorillonite in Jurassic siltstones in this study was formed from the submarine alteration of volcanic ash. The Jurassic concretion examined contained crystalline montmorillonite as the only clay mineral and the same two plagioclase species (A and B) found in Cenozoic volcanogenic sediments, thus strongly supporting a volcanic heritage for the montmorillonite (Pl. 3.2). Crystalline illite and chlorite in Jurassic sediments have mostly anhedral to subhedral mineral form (Pl. 3.3) suggesting these clays are detrital, although some could have been transformed from montmorillonitic clay during burial diagenesis. The small amounts of kaolinite in the Jurassic sediments are most likely pedogenetic as many modern studies have shown that kaolinite is a common component of soils developed on granitic and andesitic parent materials (Grim, 1953; Millot, 1971).

A review of the literature dealing with Mesozoic sediments of the study area (Kear, 1960; Hay, 1967; Fleming, 1969; Dickinson, 1971; Kear, 1971) implies that the existing clay mineralogy of the sediments is likely to be partly the result of diagenetic alterations. The Triassic-Jurassic beds of the western Hamilton and northwest Taranaki Districts (the Oparau Facies of Kear, 1971) probably represent continental shelf sediments that were deposited on the western margin of a major geosyncline bordered to the west by an active acid-intermediate volcanic arc and older foreland acid igneous and metamorphic rocks (Fleming, 1969). The poorly fossiliferous Mesozic greywackes forming the ranges of the eastern Hamilton and Taranaki Districts (the Morrinsville Facies of Kear, 1971) possibly represent continental slope sediments that were deposited in a more rapidly sinking axial zone of the geosyncline. These sediments were metamorphosed by burial to zeolite and prehnite-pumpellyite grade (see Coombs, 1960) which, to varying degrees, altered the primary mineralogy of the deposits, especially in the more deeply buried Triassic beds.

The preponderance of detrital illite and chlorite in the Mesozoic sediments is consistent with a western igneous/metamorphic provenance. However it is possible that any original montmorillonite may also have been transformed to chlorite during deep burial diagenesis through a series of mixed-layer clays (cf. Dunoyer de Segonzac, 1970, p.311). The chlorite-vermiculite and chlorite-montmorillonite in the Mesozoic sediments may then represent the intermediate stages in the transformation of montmorillonite to chlorite. The virtual absence of montmorillonite in Triassic sandstones is anticipated from the estimates of burial depths for these zeolite facies sediments, of greater than 4 km (Kear, 1960; 1971), which is in excess of the montmorillonite stability field (Burst, 1969; Dunoyer de Segonzac, 1970). Significantly, the abundance of crystalline montmorillonite in the Jurassic concretion may owe its existence to the protection from increasing pressures and circulating formational fluids by the encasing armour of calcite.

3.3.2 UPPER EOCENE TO OLIGOCENE (TE KUITI GROUP)

3.3.2A WAIKATO COAL MEASURES

The clay mineralogy of the Waikato Coal Measures is dominated by kaolinite with scarce poorly crystalline illite, heat-labile chlorite and mixed-layer clays (Figs. 3.5 and 3.6; Pl. 3.4). Clay minerals show a variety of shapes (Pl. 3.4) but subhedral kaolinite particles are most common. The sharpness of certain peaks on X-ray diffractograms suggests the kaolinitic clays are often quite crystalline.

The Waikato Coal Measures includes variably carbonaceous kaolinitic sandstones and mudstones, locally containing siderite concretions, coals and conglomerates. The vegetation associated with the carbonaceous sediments is indicative of deposition in non-marine environments, as are the siderite concretions and the rare freshwater mussels in the sediments (Sherwood, pers. comm. 1976). The one sample examined by scanning electron microscopy shows a well orientated grain fabric (Pl. 3.5). The lack of evidence for deep burial of the Waikato Coal Measures in the study area suggests the orientated grain fabric is probably the result of transportation and subsequent deposition of kaolinitic material in a non-flocculated state. This is most likely to have been achieved under freshwater conditions because kaolinite readily flocculates on entering a marine environment (Grim, 1953; Whitehouse et al., 1960; Porrenga, 1966). World wide evidence indicates that relatively intense leaching environments favour formation of kaolinite, hallyosite and sesquioxides of iron and aluminium, whereas less intense leaching conditions favour illite and montmorillonite (Jackson et al., 1948; Keller, 1962; Van der Merwe and Weber, 1963; Beaven and Dumbleton, 1966; Fieldes, 1968; Millot, 1970). Thus the climatic regime is potentially an important factor in controlling Waikato Coal Measures clay mineral assemblages. Although parent materials can exert a strong influence on soil clay mineralogy, studies in areas with a wide range of climates (e.g. Birkeland, 1969) suggest that parent materials may be of secondary importance to climate in determining clay mineral composition. To understand more fully the effect that climate may have had in forming clays in the Waikato Coal Measures it is necessary to examine the environmental conditions that existed at the time of sediment deposition. In the Eocene the South Auckland region was characterised by tectonic quiescence (Fleming, 1962). Waikato Coal Measures deposition began when a relative rise in sealevel initiated deposition on coastal plains flanked by subdued hills formed from Mesozoic sediments (Kear and Schofield, 1959; Nelson, 1973; King, 1976). The late Cretaceous to early Tertiary climate in the New Zealand region ranged from warm temperature to marginally tropical but was predominantly subtropical (Fleming, 1962; Devereux, 1967; Hornibrook, 1971; Savin, 1977). These conditions fostered dominantly chemical weathering and contemporary sediments contain widespread coal and quartoze lithologies. Under such a climate the soils developed on present day Mesozoic sandstones and mudstones are rich in kaolinite (Fieldes, 1968). Thus much of the kaolinite in Waikato Coal Measures sediments could have been detrital from soils developed on the low hills of Mesozoic basement rocks. Even if the climate was more temperate the long period of subaerial exposure of the Mesozoic rocks, from the Upper Cretaceous to Eocene (about

90 million years), might still have provided widespread pedogenetic kaolinite (cf. Birkeland, 1969).

It is well established that the kaolinite and halloysite groups of minerals are most commonly produced by leaching of metallic cations from aluminium silicates and by the addition of hydrogen ions (e.g. Grim, 1953; Millot, 1970). Given these conditions kaolinite and halloysite can form from the silicate "wreckage" or can precipitate from solution (Keller, 1968; 1970; Millot, 1970). The parent materials for Waikato Coal Measures sediments were undoubtedly Mesozoic sandstones and mudstones (Kear and Schofield, 1959) which upon weathering and subaerial erosion provided coal measure basins with detrital kaolinite and lesser poorly crystalline illite, chlorite, montmorillonite and mixed-layer clays. Further leaching of these sediments could have been accomplished in situ by acid solutions generated from sulphur compounds and/or by dilute acids generated by the passage of pore waters through the carbonaceous lithologies. The end product of these reactions would be sediments characterised by stable minerals such as quartz and kaolinite and by highly degraded clays. The abundance of quartz and kaolinite in the Waikato Coal Measures testifies to these acid leaching conditions (cf. Figs. 3.4, 3.5 and 3.6).

3.2.2B MANGAKOTUKU SILTSTONE

The clay mineralogy of the Mangakotuku Siltstone is dominated by kaolinite with smaller amounts of poorly crystalline illite, montmorillonite and mixed-layer clays and rare heat-labile chlorite (Pl. 3.6). There are, however, marked variations in the relative amounts of these minerals between different localities (Appendix Table IV.1). The sparse fauna of the Mangakotuku Siltstone comprises mainly bivalves, including thin-walled species and heavy shelled Ostrea and Eumarcia species suggesting deposition in a marginal marine environment. The abundant slightly carbonaceous mudstones of the Mangakotuku Siltstone probably accumulated in sheltered coastal lagoons characterised by generally anaerobic conditions and penecontemporaneous leaching of most carbonate material (Kear and Schofield, 1959; Nelson, 1973).

Studies of recent marine sediments (e.g. Keller, 1968; Rateev et al., 1969) show that while kaolinite is stable in the marine environment it is unlikely to form there. Thus the kaolinite is probably of terrestrial origin. Paleogeographic evidence indicates that kaolinite and poorly crystalline illites and chlorites could have been derived from the weathered Mesozoic sediments and their associated soils that bordered Mangakotuku depositional basin(s). The anhedral nature of the illite and chlorite (Pl. 3.6) suggests they are detrital. Kaolinite was probably also derived from the reworking of Waikato Coal Measures sediments (Kear and Schofield, 1959). The growth of euhedral blades of presumably authigenic kaolinite in the pore spaces of one sample (Pl. 3.7) may have been accomplished by the acid solutions which leached calcium carbonate from the sediments (cf. Keller, 1968).

The origin of the montmorillonite is more difficult to explain. The absence of volcanogenic material in the sediments would seem to rule out a volcanic heritage. It could be that changing climatic conditions across the Eocene-Oligocene boundary (Fleming, 1962; Devereux, 1967) led to the development of montmorillonite in the soils of the adjacent landmass. Some evidence for this is suggested in scanning electron micrographs (Pls. 3.8 and 3.9) wherein ball-shaped aggregates of montmorillonitic clay are of possible pedogenetic origin. While montmorillonite can form readily in the marine environment the low crystallinity of the Mangakotuku montmorillonite suggests it is more likely detrital. The similarity of clay mineral suites in concretions to those of the host mudstones (Appendix Table IV.1) rules out an early diagenetic origin for the Mangakotuku montmorillonite (cf. pp.93-94).

The marked variation in clay mineral composition between samples of Mangakotuku Siltstones from different locations (cf. Appendix Table IV.1) relates mainly to the availability of kaolinite at a particular depositional site, particularly the proximity or otherwise of the kaolinite-rich Waikato Coal Measures.

Comparison of the variable Mangakotuku clay mineral assemblages with those from the stratigraphically lower kaolinite-rich Waikato Coal Measures sediments and the stratigraphically higher montmorillonite-rich Oligocene sediments shows that the Mangakotuku deposits are mineralogically transitional between Waikato Coal Measures and younger Oligocene sediments (Figs. 3.4 to 3.6). This ties-in with previous studies (Kear and Schofield, 1959; Nelson, 1973) and personal field observations which have shown that the Te Kuiti Group represents a transgressive sequence of sediments from fresh water coal measures through to shallow marine shelf deposits (Table 3.2).

3.3.2C GLEN MASSEY SANDSTONE TO OTOROHANGA LIMESTONE

The clay fraction of Oligocene sediments is characterised by an abundance of crystalline montmorillonite (Figs. 3.5 and 3.6; Pls. 3.10 - 3.13) that is generally absent from most other Cenozoic sediments. In addition, most Oligocene sediments also contain some poorly crystalline illite and lesser amounts of poorly crystalline heat-labile chlorite and mixed-layer clays (predominantly chlorite-vermiculite and/or chlorite-montmorillonite). The montmorillonite-rich clay mineral suite persists, in general, throughout the diverse array of Oligocene lithotypes and depositional environments (Fig. 3.5).

Detailed stratigraphic and lithologic studies of sediments of the Te Kuiti Group (Kear and Schofield, 1959; Happy, 1971; Nelson, 1973) have shown that the only major source of terrigenous material for Oligocene sediments was Mesozoic sandstone and mudstone that formed a generally subdued landmass partially enclosing the marine sedimentary basins. However, the clay mineralogy of Mesozoic sediments - dominated by crystalline illite and chlorite and by mixed-layer clays - is not reflected in the Oligocene deposits (Figs. 3.5 and 3.6). Moreover, the paucity of montmorillonite in the Mesozoic sediments precludes a purely detrital origin for the abundant montmorillonite in the Oligocene sediments.

The absence of any general relationship between lithology and the frequency distribution of clay minerals in the Oligocene sediments (Figs. 3.5 and 3.6) suggests that the occurrence of montmorillonite is not a function of sediment lithology.

The clay fraction of Oligocene sediments could have evolved by one or more of the following processes: (1) detrital from Mesozoic sediments and/or from the soils developed on them; (2) neoformation or transformation in the marine environment from materials derived from the neighbouring landmasses.

To explain the possible origin of the abundant montmorillonite, it is necessary to examine in more detail the environmental conditions existing in the South Auckland region at the time of sediment deposition. Sea temperatures ranged from warm temperate in the Lower Oligocene (Lwh) to marginally subtropical in the Upper Oligocene (Lw) (Devereux, 1967; Nelson, 1973, p.561). Some floral evidence suggests the climate was perhaps drier than at most other times in the Cenozoic (Fleming, 1962, p.75). Despite some conflicting evidence as to the exact nature of the New Zealand Oligocene climate (Fleming, 1975) consensus of opinion indicates that generally warm temperate conditions prevailed (Hornibrook, 1971, Savin, 1977). The highly calcareous and fossiliferous nature of the sediments (Fig. 3.48) testifies to a dominantly biochemical sedimentary regime. Sedimentation rates were extremely low, averaging only 1-2 cm/1,000 yrs, and short periods of active sedimentation probably alternated with longer periods of slow deposition, non-deposition, and by-passing or erosion (Nelson, 1973). The small total volume of terrigenous material in the Te Kuiti Group sediments indicates regional stability of the South Auckland area during the Oligocene (Nelson, 1973, p.556).

The exact nature of the Oligocene soils is unknown but a reasonable estimate of their mineralogic composition can be made by considering the soil types developed on Mesozoic sandstones and mudstones in New Zealand at the present-day under varying climatic conditions. The yellow-brown earths constitute the most extensive zonal class of soils developed on these parent materials. Several sub-classes are recognised within the average climatic limits of yellow-brown earth formation, their clay mineralogy being related to the intensity of weathering and the severity of leaching at the soil site (Fieldes, 1968; Table 3.5). The

Weathering stage of soil	Weakly leached (50% base sat.)	Moderately leached (30-50% base sat.)	Strongly leached (30% base sat.)
Weakly weathered (6^oC; 875 - 1780 mm)	-	-	Illite* Vermiculite
Moderately weathered (11^oC; 1020-2540 mm)	Illite* Montmorillonite	Illite* Vermiculite Montmorillonite Kaolinite	Illite* Vermiculite Montmorillonite Kaolinite
Strongly weathered (14^oC; 1270-2540 mm)	Montmorillonite	Montmorillonite Kaolinite	Kaolinite

Table 3.5. The distribution of clay minerals derived from micaceous minerals in yellow-brown earths under varying weathering and leaching conditions (after Fieldes, 1968; Nelson, 1973). Illite* = illite and/or interlayered hydrous micas. Leaching defined on the percentage base saturation shown.

evolution of montmorillonite is favoured under warmer climates and poorly drained conditions where metallic ions and silica released during hydrolysis can accumulate (Keller, 1970), as is the case for several modern calcimorphic soils (Millot, 1970, p.111 and 325). Any growth of montmorillonite in the soils would enhance the development of a compact, tight and therefore poorly drained soil under wetting and drying conditions, and thus promote conditions conducive to further montmorillonite development, particularly by neoformation. Such conditions undoubtedly existed locally on the adjacent low lying, stable landmass.

Primary clay minerals of the Mesozoic sandstones and mudstones, plus secondary sericite and chlorite produced by weathering of feldspar and ferromagnesian minerals, may be transformed by the following degradational pathways to montmorillonite (Millot, 1970, p.307):

$$I \longrightarrow I\text{-}V \longrightarrow V \longrightarrow V\text{-}M \longrightarrow M$$

$$C \longrightarrow C\text{-}V \longrightarrow V \longrightarrow V\text{-}M \longrightarrow M$$

More intensive leaching may yield kaolinite and secondary silica. Treatment of selected Oligocene samples with KOH (Table 2.6) indicates that the montmorillonites are unlikely to have originated from micaceous material. Any transformed montmorillonite in the soil horizons developed on Mesozoic sediments and subsequently transported to the Oligocene marine sedimentary basins would be stable within this new environment (Millot, 1970, p.361). However, if pedogenically transformed montmorillonite was a major component of the Oligocene sediments one might have expected to find more clay minerals intermediate in the soil transformation sequence to montmorillonite (cf. Fig. 2.6). It is concluded that detrital pedogenic montmorillonite is only a minor component in the Oligocene clay mineral suite but that the small amounts of anhedral, poorly crystalline illite, chlorite and mixed-layer clays are most probably derived from the Oligocene soils (Pls. 3.11 and 3.12).

Detailed studies (Nelson, 1973) have shown that during the Oligocene, the adjacent tectonically stable landmass was a source of dissolved chemical products for the marine sedimentary basins and that sedimentation in the shallow carbonate-rich seas was extremely slow. Under conditions such as these, transformation to and neoformation of montmorillonitic clay minerals is especially favoured (Millot, 1970). Of the two processes the writer prefers neoformation as playing the major role in the development of Oligocene montmorillonites for the following reasons: (1) The general paucity of mixed-layer clay minerals (Figs. 2.6 and 3.5) which expectedly would be more common if the montmorillonites were formed primarily by transformation of other clay minerals. (2) The occurrence of lath-shaped aggregates of montmorillonites (Pls. 3.12 and 3.13) is typical of neoformed montmorillonite (Roy and Sand, 1956; Kotelnikov, 1965). (3) The extremely fine grain-size of the montmorillonite (Pls. 3.11 - 3.13) is suggestive of progressive development from collodial material, as demonstrated by laboratory synthesis (Beutelspacher and Van der Marel, 1968; Millot, 1970, pp.336-354). (4) The occurrence of low-temperature cristobalite in the Oligocene sediments (Pls. 3.11 - 3.13; Appendix Tables III.1 and IV.1) supports neoformation (cf. Jeans, 1968, p.323; Millot, 1970, p.201). (5) The consistently high crystallinity of the montmorillonite (Fig. 3.6) is considered to be more indicative of neoformation which gives rise in the first instance to a well defined crystal structure. Transformation, on the other hand, which involves the reorganisation of existing crystal lattices, might be expected to produce a range of crystallinities.

Studies of the fixation of ions by clay minerals (e.g. Powers, 1957, 1959; Carroll and Starkey, 1960; Harder, 1972; Sayles and Mangelsdorf, 1977) have shown that the absorption of Mg^{2+} ions, a prerequisite for the formation of Mg-montmorillonite, is much greater at or near the sediment surface than at depth, where absorption of K^+ and the development of illite prevails. Several studies (e.g. Burst, 1959; Dunoyer de Segonzac, 1970) have shown that illite and also chlorite increase at the expense of montmorillonite during diagenesis. It is probable, therefore, that the Mg-montmorillonite of the Oligocene sediments developed at or near the sediment surface. Assuming that material transported to the sedimentary basins contains all possible products of chemical weathering then depending on the degree of saturation of ions in the marine environment, the degraded minerals will begin aggradational transformations, with faster rates under more saturated conditions. Studies on clay mineral synthesis by Millot (1970, pp.323-347) and Harder (1972) have shown that, if conditions are favourable, amorphous substances and solutions will take part in neoformations of new clay minerals. Both workers emphasised the important role that the chemistry of the environment plays in the neoformation of clay minerals. For montmorillonite to have been neoformed during shallow burial a magnesium-rich environment is a prerequisite (Harder, 1972). This magnesium may have been supplied from sea water supplemented by the incongruent dissolution of skeletal magnesium-calcite close below the sea floor (Beu et al., 1972, p.279). Anaerobic conditions existing in the upper parts of the shallow burial zone may increase the pH to over 9. Experiments have shown that montmorillonite precipitation is enhanced by increasing pH, and in fact at 20^{o}C and pH9 a Mg^{2+} concentration of 30ppm in the pore waters (cf. "average seawater" Mg^{2+} concentration approximates 1000 ppm) may initiate montmorillonite neoformation (Harder, 1972).

In many clay mineral studies, the presence of large quantities of montmorillonite has been attributed directly to the alteration of volcanic material. The complete absence of primary volcanic material in Oligocene sediments of the South Auckland region (Nelson, 1973), together with the absence of beds enriched in trioctahedral montmorillonite, analcite and plagioclase of both A and B species, minerals found in this study to be indicative of volcanism (see p.78), probably rules out a volcanic heritage for the Oligocene montmorillonite. Despite the absence of mineralogic evidence, however, a possible volcanic influence exists if a paleogeographic reconstruction for New Zealand in the Oligocene is based on the tenet that transcurrent movement on the Alpine Fault had not yet begun (Carter and Norris, 1976; Ballance, 1976; Nelson and Hume, 1977). Such a reconstruction (e.g. Griffiths and Varne, 1972; Griffiths, 1973; Wellman, 1973; Carter, 1974) shows that the Oligocene sedimentary basins of the study area lay some 4^{o} of latitude south of their present position relative to basement rocks on the eastern side of the Alpine Fault. The Oligocene sediments could therefore have had as their source Mesozoic sandstones and mudstones now exposed in the northern South Island, in the southern North Island axial ranges, and in the Herangi Range. Several small areas of volcanic rocks outcrop in the calcareous sandstones and mudstones of the Oligocene Esk Formation (Gregg, 1964) in the northeast of the South Island. It is remotely possible that these volcanics were a source of montmorillonite in the South Auckland Oligocene sediments, and that the subsequent Miocene illitic and chloritic clay mineral assemblages reflect the cessation of this volcanism and/or a relative northward movement of the sedimentary basins away from this source by dextral movement along the Alpine Fault. However the writer considers that the volcanism associated with the Esk Formation probably played no part in the development of the montmorillonite-rich Oligocene clay fraction because, among other reasons, the Duntroonian (Ld) age suggested for the Esk volcanics (Gregg, 1964) means they could not explain the abundant crystalline montmorillonite in the Whaingaroa Siltstone of Whaingaroan (Lwh) age.

The Oligocene was a period of general tectonic quiescence and low relief, and vegetation flourished under a warm temperate to subtropical climate (Fleming, 1962; Hornibrook, 1971). Under these conditions leaching prevails in the soil profile and the major constituents reaching the sedimentary basins are dissolved ions (e.g. Ca, Mg and Si), the most stable minerals (e.g. quartz) and the residues of clay minerals (highly degraded illite and chlorite). These dissolved constituents recombined to form marine chemical sediments dominated by carbonates, crystalline montmorillonite, and cherts (α-cristobalite) (cf. Figs. 3.4 to 3.6).

Although montmorillonite is the dominant clay mineral in most Oligocene sediments there are some exceptions. Limestones and sandstones of on-basement facies expectedly show a significant increase in the abundance of detrital kaolinite, illite and chlorite at the expense of montmorillonite. Similarly sediments overlying coal measures (e.g. Mangapehi sandstones and Mangakotuku siltstones) show an increase in kaolinite and quartz, most probably as a result of reworking of the coal measure deposits. The relatively high content of kaolinite in sediments from the Awakino area (especially locality 28) is probably a function of their proximity to the more tectonically active Herangi Range (Cope and Reed, 1967; Nelson, 1973; Nelson and Hume, 1977). The banded Aotea sandstone in the Awakino area contains abundant kaolinite and very poorly crystalline illite and chlorite. Comparison with the clay mineral assemblages of massive Aotea sandstones (i.e. montmorillonite-rich sandstones), suggests that the kaolinite has formed by the acid leaching of montmorillonitic and chloritic material in sandstones of "normal" montmorillonite-rich composition. This supports the contention that the banded Aotea sandstone ".... results mainly from active leaching of calcium carbonate (mainly cement) from outcrops at the present day" (Nelson, 1973, p.514). Some samples containing glauconitic material show an increase in illite abundance at the expense of montmorillonite in the clay fraction. Such sediments generally occur as intensely burrowed basal facies (e.g. basal Aotea Sandstone lithofacies). Nelson (1973, p.462) suggested that the low sedimentation rates and reducing conditions occurring in these facies may have favoured the uptake of K^+ and Fe^{2+} into montmorillonitic lattices thereby promoting their transformation to glauconite.

3.3.3 LOWER MIOCENE (MAHOENUI AND MOKAU GROUPS)

The clay mineralogy of Lower Miocene sediments is characterised by montmorillonite and crystalline illite, with minor mixed-layer clays and crystalline chlorite (Figs. 3.5 and 3.6; Pls. 3.14 to 3.17). Abundant kaolinite is restricted to the Maryville Coal Measures. Compared to the Oligocene sediments, the Lower Miocene deposits contain less montmorillonite, which is also of lower crystallinity, and show a marked increase in both illite and chlorite abundance and crystallinity (Figs. 3.5 and 3.6), together with the appearance of crystalline chlorite (see pp.44-51 and Fig. 3.7), a mineral species not present in Oligocene sediments. Through the Lower Miocene sediments there is a systematic decrease in the abundance of montmorillonite and an increase in the content of crystalline illite and chlorite and of mixed-layer clays. These changes persist through all size fractions (Fig. 2.6).

Possible origins for the Lower Miocene clays include their being detrital from sediments and soils on the adjacent

Lower Miocene landmass(es), or transformed and neoformed in the marine sedimentary basins and/or during sediment diagenesis. This section considers these possible origins in turn.

In Lower Miocene sediments illite and chlorite particles are generally anhedral and the illite commonly shows irregular diffraction patterns produced by strain within the lattice (Pls. 3.14 and 3.16). This latter phenomenon is commonly associated with clay material that has undergone advanced weathering (cf. p. 56). These features strongly suggest that the majority of illite and chlorite in Lower Miocene sediments is detrital. Both scanning electron microscopy (Pls. 3.15 and 3.17) and thin section examination show no evidence for the authigenic growth of clay minerals in samples. Treatment of selected samples with KOH (Table 2.6) indicates that the montmorillonites are unlikely to have originated via the transformation of micaceous material, either in the soils of the source area or during subsequent sediment diagenesis. Thin section examination indicates that the high concentration of clay minerals in bulk samples (Fig. 3.48) is caused partly by increased amounts of illite and chlorite in the silt and sand size fractions. The decrease in the content of calcium carbonate (Fig. 3.48) coupled with the increase in abundance of rock fragments (Fig. 3.17), on passing from the Oligocene into and through the Lower Miocene sediments testify to a change from a predominantly biochemical to a more detrital sedimentary regime. From thin section data, the increase in plagioclase feldspar content shown by bulk mineralogic analysis reflects mainly an increase in rock fragment abundance (cf. Figs. 3.4A and 3.17). If Mahoenui Group clays are detrital from hinterland sediments, they would have as their source Mesozoic sandstones and mudstones and possibly Oligocene sedimentary rocks that probably flanked the north-south trending Lower Miocene sedimentary basins (Fig. 1.3). Data from this study and others (e.g. Fieldes et al., 1968; Nelson, 1973) show that direct erosion and breakdown of Mesozoic sandstones and mudstones would be expected to produce a clay mineral assemblage rich in crystalline illite and chlorite and in mixed-layer clay minerals. While these clay minerals are present in the Mahoenui clay mineral suite, they are in general less abundant than montmorillonite. Thin section analysis of the Mahoenui sandstones reveals the presence of sandstone and mudstone rock fragments and flakes of sand size illite and chlorite compatible with a Mesozoic source for the sediments. Sand/shale ratios determined from oil well logs (McQuillan, 1977) suggest that during Mahoenui deposition, the present day Mesozoic basement outcrops may have been sufficiently elevated to provide clastic material from both the eastern and western margins of the north-south trending Mahoenui Basin. The abundant montmorillonite in the Mahoenui sediments cannot, however, have been derived from direct erosion of Mesozoic sediments (Fig. 3.5). The present areal distribution of Te Kuiti Group sediments shows they probably cropped out to the west, north, and east of Mahoenui sedimentary basins (Fig. 1.3). Furthermore, oil well data demonstrate that Oligocene calcareous sediments, presumably montmorillonite-rich, were also more extensive to the southeast (McQuillan, 1977). Direct erosion of these Te Kuiti Group lithologies would have supplied crystalline montmorillonite to the Mahoenui sediments (Figs. 3.5 and 3.6) and would possibly also account for the locally common occurrence of calcium carbonate in the same sediments (Appendix Table III.1).

Thus contemporaneous erosion of the Mesozoic and Te Kuiti sediments could explain the clay mineral assemblages in the Mahoenui sediments. The less crystalline nature of the montmorillonite in the Lower Miocene mudstones compared to that in the Oligocene mudstones is considered to reflect degradation of the montmorillonite during pedogenesis and transport to the marine sedimentary basins. Moreover the lowered crystallinity would suggest that the pedogenetic development of montmorillonite on the adjacent landmass was not favoured.

In the central northern parts of the North Wanganui Basin the Taumatamaire calcareous mudstones were deposited in water depths greater than about 100 metres (Happy, 1971, p.104). Except for the presence of crystalline illite and some chlorite in these mudstones their clay mineralogy is very similar to that of the Oligocene mudstones, being dominated by moderately crystalline montmorillonite, derived most probably from Oligocene sediments. The great thickness of sediment in the vicinity of Ohura, where 1384 m of Mahoenui sediment is recorded in one oil well log, is considered by Glennie (1959) to be the result of an eastward downfaulting along the Ohura Fault.

In the western North Wanganui Basin adjacent to the Herangi Range, the Taumatamaire sandy mudstones and interbedded sandstones and limestones were deposited in water depths shallower than 100 metres (Happy, 1971, p.104). Compared to the more central mudstones, these sediments contain larger proportions of illite and chlorite relative to montmorillonite. In addition the illite and chlorite is more crystalline, the sediments contain kaolinite, and the mudstones are typically more sandy and contain more quartz and feldspar relative to calcite. These characteristics can be attributed to the adjacent Herangi Range being uplifted during deposition of the Taumatamaire sediments (Happy, 1971, p.118; Nelson, 1973). The slightly increased quantities of montmorillonite in the limestone units of the Taumatamaire Formation (Appendix Table IV.1) probably reflects the deposition of these units during periods of minimal influx of detrital (illite and chlorite) material.

In the eastern North Wanganui Basin, Mahoenui flysch-type sediments (Taumarunui Formation) were deposited in deep waters in a north-south trending basin (Glennie, 1959). The great thickness of these sediments (>600m), their relatively high sedimentation rate (Nelson and Hume, 1977), the narrowness of the trough, and its alignment with major faults and basement highs in the regions, suggest that deposition and basin subsidence were primarily controlled by contemporaneous tectonics. Glennie (1959) suggested that the sediment was derived from a major Lower Miocene river mouth to the north. The clay fraction of Taumarunui sediments shows relatively high concentrations of kaolinite and illite in the north as compared to the south, where montmorillonite dominates (Fig. 3.10). Similar lateral variations in clay mineral abundances have been used by a number of writers to indicate directions of sediment transport (e.g. Shover, 1964; Parham, 1966; Porrenga, 1966), and it is possible that clay mineral trends in the Taumarunui sediments may be used likewise. Because of this it is pertinent to review the factors that can cause such variations in sediments.

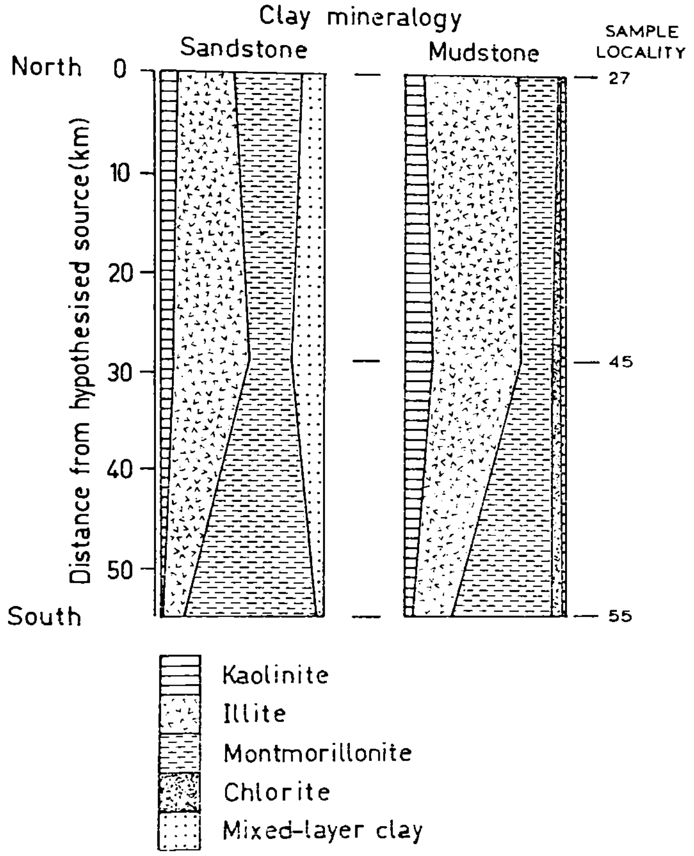


Fig. 3.10. Lateral clay mineral distributions in the flysch-type Taumarunui Formation.

Systematic trends in clay mineral abundance in sediments have been attributed to the processes of chemical alteration, differential flocculation and size segregation. The chemical alteration viewpoint considers that, upon entering the marine environment, those clay minerals not in chemical equilibrium with seawater will alter, via transformations, to more stable forms. Johns and Grim (1958) attributed clay mineralogic changes in Mississippi River delta sediments to transformation of part of the montmorillonite fraction to illite by K^+ uptake within the montmorillonite lattice. Milne and Early (1958) considered that similar factors determined clay mineral distribution patterns in Gulf of Mexico sediments. Powers (1957) recorded the development of chlorite from illite via mixed-layer clay mineral intermediates in Chesapeake Bay sediments. Differential flocculation can occur when clay particles are transported from a non-marine to a marine environment. Clay minerals normally have a negative surface charge arising from broken intermolecular bonds at the edges of the clay mineral lattice and from cationic substitutions within the lattice. The total negative charge is balanced by a double layer of hydrated cations. The stability of this layer is dependent on the ionic concentration of the surrounding water, and as the ionic concentration rises (e.g. on moving from a non-marine to a marine environment), there is a tendency for clay particles to flocculate, the readiness with which clay minerals flocculate depends on a complex array of factors including ionic concentration, suspended sediment concentration, water turbulence, water temperature and the concentration of various carbohydrate and organic complexes in the medium (Weaver, 1958a; Whitehouse <u>et al</u>., 1960; Gibbs, 1977). In general illite, chlorite and kaolinite flocculate before montmorillonite. Gibbs (1977) considered that particle size segregation and differential settling were responsible for the decreased abundance of montmorillonite but increased abundance of illite and kaolinite in Atlantic Ocean sediments as the Amazon River source was approached. Furthermore, he contended that preferential flocculation may be unimportant in the marine environment where natural organic and metalloid coatings give all clay minerals similar flocculation properties.

Comparison of the above works with results from this study (Fig. 3.10) suggests that lateral variations in clay mineral abundance in Taumarunui sediments can be attributed to either preferential flocculation or sedimentation of clay materials. These explanations would support Glennie's (1959) argument for a northern source for the Taumarunui flysch-type sediments. The sampling density in Taumarunui sediments was too low to confirm the relative importance of these processes in determining clay mineral distributions. Furthermore, paleogeographic and petrologic data suggest it is probable that some of the montmorillonite in the flysch sediments was derived by the slumping of montmorillonite-rich Taumatamaire shelf sediments on the western flanks of the trough, and that some of the illite, chlorite and mixed-layer clays were derived from erosion of the uplifted Mesozoic ranges in the east. Such multiple sources could account for the variable clay mineralogy of the Taumarunui sediments (cf. Appendix Table IV.1).

The Mahoenui sediments are dominated by mudstones. Evidence already presented suggests significant relief and rapid mechanical erosion were characteristic of the early Lower Miocene (Nelson and Hume, 1977). Under such conditions, the Mahoenui Group would be expected to contain an abundance of coarser terrigenous material. This apparent ambiguity is considered to be largely provenance-controlled. In the western part of the study area, Mesozoic sediments are largely relatively incompetant Jurassic shelf mudstones and muddy sandstones (Kear, 1971). These sediments were deeply weathered during the Cretaceous and early Tertiary (see pp.99-100), with thick soil development (Fieldes _et al_., 1968). Oligocene sediments are largely biochemical products with the terrigenous fraction being dominated by argillaceous products. Earliest Miocene tectonism would have resulted in the stripping of these Mesozoic and Oligocene sediments as fault blocks were elevated. The abundant terrigenous muds probably accumulated rapidly in both nearshore and offshore environments (cf. McCave, 1971).

Possible sources for detrital Mokau Group clays include the same Mesozoic sandstones and mudstones and Oligocene sediments that were sources for the Mahoenui clays. In Mokau times, the restriction of sediments to that area west of the Ohura Fault and the rapid eastward diminution in thickness, accompanied by increased incidence of conglomeratic intervals as the Ohura Fault is approached, provide evidence of the control of Mokau sedimentation by renewed movements on the Ohura Fault. McQuillan (1977) noted that the eastern downthrow on the Ohura Fault in the late Oligocene - early Lower Miocene, which gave rise to the thick Mahoenui sequence, was reversed in the late Lower Miocene such that sedimentary thinning, and erosion became the mode. It is likely that there was considerable reworking of Mahoenui sediments because of uplift of the area east of Taumarunui. Stainton and Gibson (1964) determined that upwards of 300 m of section was eroded from the Mahoenui Group over the area east of Ohura before the Mokau sediments were deposited.

Thick Lower Mokau sandstones, with clay mineral assemblages characterised by montmorillonite, crystalline illite, chlorite and mixed-layer clays, were deposited under shallow, nearshore conditions. The montmorillonite in the sandstones is probably mainly detrital from the readily erodible shelf and flysch-type Mahoenui sediments and perhaps also Oligocene mudstones that were elevated both to the east of the depositional basin by movement on the Ohura Fault and to the north by movement in the vicinity of the Piopio Threshold (Fig. 3.3). The lower crystallinity of the montmorillonite in the Mokau sediments compared to that in the source Mahoenui sediments is considered to reflect minor degradation of the montmorillonite in the soil profile. The Mahoenui and Oligocene mudstones east of the Ohura Fault probably flanked the western margins of high Mesozoic basement ranges. The headwaters of rivers traversing the exposed Mahoenui and Oligocene sediments probably tapped these Mesozoic sediments which supplied crystalline illite and chlorite and mixed-layer clays to the Mokau sedimentary basins. Landmasses to the north and west probably supplied similar detritus.

The clay fraction of Maryville Coal Measures sediments contains abundant kaolinite and poorly crystalline illite, scarce mixed-layer clays and poorly crystalline chlorite and montmorillonite (Appendix Table IV.1). The bulk samples contain quartz and a high content of kaolinitic clay minerals (Appendix Table III.1). The kaolinitic sandstones and mudstones are associated with abundant carbonaceous material, including coal at certain horizons. The Maryville Coal Measures are considered to have been deposited when paralic conditions developed on wide plains built up of Lower Miocene sediments (Henderson and Ongly, 1923, p.72). The clay fraction of Maryville Coal Measures sediments is unlikely to be detrital from pre-Mohakatino lithologies for, with the exception of the lowest Te Kuiti Group sediments, these lithologies are kaolinite-poor. Maryville Coal Measures sediments are strikingly similar in composition and occurrence to Waikato Coal Measures sediments, suggesting a generally similar origin. The kaolinite in the Waikato Coal Measures is considered to be in part detrital from soils developed on Mesozoic sediments, and in part transformed from sediments that underwent acid leaching generated by carbonaceous material (cf. pp. 97-100). It is unlikely that Maryville Coal Measures clays are predominantly detrital from soils developed on the adjacent landmasses as, compared to Waikato Coal Measures times, the more temperate climatic conditions (Devereux, 1967; Hornibrook, 1971) and the shorter period of time (cf. Birkeland, 1969) would have been much less conducive to kaolinite genesis.

The predominance of quartz and kaolinite associated with abundant carbonaceous material is strongly suggestive of kaolinite formation by in situ leaching associated with the passage of acid pore waters through the sediments (cf. Keller, 1970, pp.797-799).

In Upper Mokau times predominantly deep water muds were deposited. The source of the Upper Mokau clays is considered to be similar to that described for the Lower Mokau sandstones. However, the higher proportions of illite and chlorite relative to montmorillonite in the Upper Mokau sediments suggests that the Mesozoic sandstone and mudstone provenance was more dominant.

The Lower Miocene climate was apparently mainly subtropical (Devereux, 1967; Hornibrook, 1971; Savin, 1977), with some floral evidence suggesting a higher rainfall compared to the Oligocene (Fleming, 1962). There is considerable evidence to indicate that in contrast to the Oligocene the Lower Miocene was a period of more active tectonism (Ballance, 1976; Nelson and Hume, 1977). These data, together with mineralogic evidence, suggest that mechanical erosion of uplands would be pronounced, soil development would be reduced. Clay minerals weathered from parent rocks would have mainly by-passed the soil profile and have been rapidly flushed to the marine basins. The increasing abundance of crystalline illite and chlorite, and also of sandstone and mudstone rock fragments in Mahoenui through to Mokau sediments, suggests that the Oligocene sediment source became less important relative to the Mesozoic sandstone and mudstone source as time proceeded.

There is little evidence to support an authigenic origin for the clay minerals in the Lower Miocene sediments. If the minerals were of transformational origin, larger quantities of mixed-layer clays would be expected (Fig. 2.6). The Lower Miocene montmorillonites are unlikely to be neoformed as the clay fraction contains no lath-shaped particles or aggregates of montmorillonite, no α-cristobalite and the montmorillonite is of only moderate crystallinity (cf. p.106). Also, environmental conditions of increased tectonism, high relief, rapid erosion and high rates of sedimentation are not conducive to the development of transformed and neoformed clays (cf. Millot, 1970). The absence of primary volcanic deposits and volcanogenic material (Table 3.6) in the sediments precludes a volcanic heritage for the clay mineral assemblages.

The similarity of clay mineral assemblages between concretions and their host mudstones is taken to indicate that the Lower Miocene clays are unlikely to have formed during burial diagenesis (cf.pp. 93-94).

3.3.4 MIDDLE MIOCENE (MOHAKATINO GROUP)

Middle Miocene sediments are considered to have accumulated slowly in a variety of shallow subtidal environments (Hay, 1967; Happy, 1971). They consist of a variety of lithologies (Table 3.3) displaying variable clay mineral assemblages (Figs. 3.5 and 3.6). The sediments are unique in as much as they contain much andesitic tuffaceous material, the source of which is inferred from magnetic anomalies to have been a volcanic complex lying some 40 km offshore from Awakino (Hay, 1967; Hatherton, 1968).

The tuffaceous Mangarara, Purupuru and Omoao Formations have distinctive grain fabrics (Pls. 3.18 - 3.20) and mineralogic compositions (Pl. 3.21; Table 3.6) characterised by high plagioclase to quartz ratios and by the presence of α-cristobalite cement (Pls. 3.18 and 3.19), analcite and amphiboles (Table 3.6). The clay fractions contain abundant montmorillonite (Fig. 3.5; Pl. 3.21). The mineralogy of tuffaceous sandstones contrasts markedly with that of interbedded mudstones and suggests a fundamental difference in origin. The volcanogenic character of the sandstones is indicated by their bulk mineralogy (Table 3.6) and by the abundance of andesitic rock fragments in the

	Sandstone	**Mudstone**
Bulk Sample	**Plagioclase > Clay minerals > Quartz** **Plagioclase A ≈ Plagioclase B** **Characteristic accessory minerals** **- α-cristobalite** **- Analcite** **- Amphibole**	**Clay minerals > Plagioclase > Quartz** **Plagioclase A ≫ Plagioclase B** **Accessory minerals sometimes present** **- α-cristobalite** **- Amphibole**
Clay Fraction $< 2\mu$	**Illite** **- scarce** **- low crystallinity** **Montmorillonite** **- abundant** **- high crystallinity** **- trioctahedral**	**Illite** **- abundant** **- high crystallinity** **Montmorillonite** **- scarce** **- low crystallinity** **- dioctahedral**

Table 3.6. Comparison of the mineralogic properties of volcanogenic sandstones and mudstones as determined by XRD analysis.

sediments (Fig. 3.17). Fabric analysis and thin section examination show euhedral feldspars and clay material cemented by α-cristobalite (Pls. 3.18 and 3.19), and both the bulk mineral and clay size fraction contain analcite.

Tuffaceous sandstones contain very abundant crystalline, iron-rich, trioctahedral montmorillonite (Table 3.6). In sediments examined in this study, this mineralogy is unique to volcanogenic sandstones. If the clays in the sandstones were detrital from the Mesozoic to Lower Miocene sediments, one might have expected them to contain dioctahedral montmorillonite, crystalline illite and chlorite and mixed-layer clays (cf. Figs. 3.5 and 3.6). There are several features that suggest the trioctahedral montmorillonite is authigenic. The montmorillonite is highly crystalline (Fig. 3.6), lath-shaped (Pl. 3.21 cf. Pls. 3.11 - 3.13), ubiquitous in some samples (cf. Almon et al., 1976; Wilson and Pittman, 1977), and is also associated with α-cristobalite (Pl. 3.18; Jeans, 1968) and analcite (cf. Deffeyes, 1959; Deer, et al., 1967). Many writers (e.g. Grim, 1953; Millot, 1970; Almon et al., 1976) have described the development of montmorillonite from glassy volcanic material in the marine environment. It appears that volcanic glass may be improverished in cations until the disordered lattice attains a state favourable to the growth of montmorillonite crystals. Millot (1970) described this process as neoformation by subtraction. This chemical reorganisation takes place so readily in the marine environment that volcanic glass may alter to montmorillonite even during sediment transport, and alteration is certainly well advanced by early diagenesis (Dunoyer de Segonzac, 1970, p.293). Mineralogic data suggests that montmorillonite formation took place prior to α-cristobalite formation (Pl. 3.19) and during, or soon after, sediment deposition, as α-cristobalite is normally early diagenetic in origin (Iijima and Utada, 1966, p.337; Misutani, 1970). In sediments where alteration takes place in situ montmorillonite may be pseudomorphic after glass shards (e.g. Coombs, 1960); no evidence for this is seen in the Mohakatino sediments. A literature search produced two other studies (Aoyagi, 1967; Almon et al., 1976) in which tuffaceous sandstones show distinctive clay fraction mineralogies. In both these studies the montmorillonite was considered to be authigenic and to have formed from glassy volcanic material in the marine environment; however, neither study distinguished the montmorillonite as dioctahedral or trioctahedral. Aoyagi's (1967) work is poorly documented, but Almon et al. (1976) presented chemical data which suggested that the Fe/Mg-rich montmorillonite was neoformed and reflected the chemistry of early stage pore fluids. It is therefore possible that the Fe/Mg-rich trioctahedral crystalline montmorillonite of the Mohakatino tuffaceous sandstones were the product of the early diagenetic alteration of glassy volcanic material in the marine environment, perhaps originating from volcanic airfall detritus. The illite, chlorite and montmorillonite in the larger size fractions of the same samples (Fig. 2.6) are probably detrital from the erosion of onshore pre-Mohakatino sediments.

Petrographic data suggests that the Mangarara calcareous sandstones and rare limestones had a mixed provenance that included pre-Middle Miocene sediments and Mohakatino volcanogenic products. Compared to the Mangarara and Purupuru sandstones, the clay fraction of the overlying Omoao sediments contains smaller amounts of crystalline montmorillonite and greater quantities of more crystalline illite and chlorite. While the Purupuru Formation was formed during a period of intense volcanism, the Omoao sediments represent products derived mainly from the erosion of pre-Mohakatino Tertiary sediments intermixed with pyroclastic material supplied during the dying phases of volcanic activity. The massive, porous Omoao sandstone (i.e. Om: 10303) contains analcite which is suggestive, at least in part, of a volcanic origin, but unexpectedly the clay fraction is dominated by illite and chlorite. The permeable nature of this sandstone, and the low crystallinity of the illite and chlorite, suggest that extensive leaching has taken place so that any montmorillonite originally present would have been degraded to illite.

The more volcanogenic character (cf. Table 3.6) of western Mangarara and Purupuru sandstones supports the western source proposed for the volcanics by Glennie (1957), Hay (1967) and Hatherton (1968).

The clay mineralogy of tuffaceous mudstones differs markedly from that of tuffaceous sandstones (Table 3.6), suggesting a different origin for the clay minerals. Crystalline illite is dominant, and crystalline chlorite, dioctahedral montmorillonite of medium crystallinity and mixed-layer clays are subequal in abundance (Table 3.6; Figs. 3.5 and 3.6). The mudstone fabric (Pl. 3.20), and the abundance of anhedral illite and chlorite, suggests these minerals are detrital. Together with montmorillonite (and mixed-layer clays) they were probably derived from Mesozoic to Lower Miocene sediments in the catchments surrounding the middle Miocene depocentres. The numerous lath-shaped dioctahedral montmorillonite particles (Pl. 3.21) and the presence of α-cristobalite may indicate the montmorillonite is neoformed (cf. p.107). However, the mudstone montmorillonite is notably less crystalline compared to the trioctahedral montmorillonite of the tuffaceous sandstones. Thus montmorillonite in the tuffaceous mudstones is perhaps polygenetic, originating from the reworking of both subaerially exposed Mohakatino tephras and pre-Mohakatino non-volcanic sediments. The lower crystallinity of the montmorillonite in the mudstones compared to the sandstones perhaps reflects a mixing of crystalline neoformed volcanogenic montmorillonite and less crystalline detrital montmorillonite. Furthermore treatment with KOH (Table 2.6) suggests the montmorillonite is unlikely to have been transformed from micaceous material.

The younger members of the Mohakatino Group (Ferry and Tawariki Mudstones) are non-tuffaceous (Table 3.3). In general their clay fraction is similar to that of the earlier Mokau sediments, although there is an increase in the abundance and crystallinity of illite and chlorite (Figs. 3.5 and 3.6). A detrital origin for the illite and chlorite is supported by their anhedral form and by clumps or books of clay minerals between grains of quartz, feldspar and rock fragments (Pls. 3.22 and 3.23). The abundance of sandstone and mudstone rock fragments in the sandstones (Fig. 3.17) suggests that Mesozoic sediments were a possible source for the crystalline illite and chlorite. Montmorillonite is unlikely to have been formed by the transformation of micaceous material (cf. Table 2.6). Montmorillonite, mixed-layer clays, illite and chlorite could have been reworked from Oligocene to Lower Miocene sediments. The high crystallinity of detrital chlorite suggests little degradation of this clay by pedogenesis of either Mesozoic or Tertiary source sediments.

Variations in sand/shale ratios indicate that land masses east and west of the basin supplied clastic sediments, particularly from the northeast and from the Herangi Range area in the northwest (McQuillan, 1977). Intraformational slump structures in Middle Miocene sediments adjacent to the Patea-Tongaporutu Gravity High (Henderson and Ongley, 1923) suggest local tectonic movements throughout this period, which probably aided erosion and reworking of Mesozoic to Lower Miocene sediments.

3.3.5 UPPER MIOCENE TO PLIOCENE (MT. MESSENGER SANDSTONE, URENUI SILTSTONE, MATEMATEAONGA SANDSTONE AND TANGAHOE MUDSTONE)

The clay mineralogy of Upper Miocene to Pliocene sediments is characterised by abundant crystalline illite, common poorly crystalline montmorillonite and crystalline chlorite, scarce mixed-layer clays and rare kaolinite (Figs. 3.5 and 3.6). With decreasing age the sediments generally show an increasing amount of montmorillonite and a decreasing content of illite and chlorite (Fig. 3.5).

Possible origins for the Upper Miocene to Pliocene clays include their being detrital from sediments and soils on the adjacent landmass; transformed or neoformed in the marine sedimentary basins and/or during sediment diagenesis.

A detrital origin for the bulk of the Upper Miocene to Pliocene clays is suggested by the dominance of anhedral illite and chlorite (Pls. 3.24, 3.26, 3.28 and 3.31) and by certain grain fabric relationships which show clay material coating, and in packets between, sand and silt-size minerals or rock fragments (Pls. 3.25, 3.29, 3.30 and 3.32). The montmorillonite in these samples is unlikely to be neoformed as it is of low crystallinity (Fig. 3.6) and lath-shaped particles are rare. Furthermore the montmorillonite of Upper Miocene to Pliocene clays is unlikely to have been transformed from micaceous material (cf. Fig. 2.6). Scanning electron micrograph and thin section data suggest that the abundance of clay minerals in bulk sediments (Fig. 3.4B) is due to the common occurrence of coarse-grained detrital micas and chlorite in the sediments (Pl. 3.32) as well as the relatively high content of sedimentary and volcanic rock fragments in sandstones (Fig. 3.17).

In the Upper Miocene, sediment was supplied to the North Wanganui Basin from terrigenous sources to the south and west, while tuffaceous material continued to be supplied from volcanics centred offshore from the present Taranaki coast (Hay, 1967; Hatherton, 1968; McQuillan, 1977). To the west, strong uplift and/or flexing of the Patea-Tongaporutu High at the beginning of Mt. Messenger times is indicated by spectacular slump structures and conglomerate-filled channels in the predominantly sandy Mt. Messenger sediments (Glennie, 1958). To the south there is a general coarsening of sediments towards the basement Pipiriki High which persisted as the southern margin to the North Wanganui Basin (Figs. 3.1 and 3.3). The northern boundary of the Basin is unknown beyond the present limit of erosion (Fig. 1.3). Urenui times saw the deposition of predominantly mudstones and a decrease of activity of the Patea-Tongaporutu High which was eventually buried, although the Pipiriki High persisted as a southern barrier (McQuillan, 1977). Rare tuffaceous beds in the Urenui sequence (Pl. 1.21) indicate that volcanism continued at this time. As the depositional basins shallowed, mudstones were replaced by more sandy, gravelly and shelly facies of the Matemateaonga Formation (Pl. 3.30). This change in sedimentary pattern coincided with a shift of depositional centres to the south and east as a result of major tilting hinged along an east-west axis approximately coincident with the Pipiriki High (McQuillan, 1977). A continuation of this trend led to the deposition of predominantly deep water muds in Tangahoe times in the southern North and northern South Wanganui Basins.

It is possible that much of the crystalline illite and chlorite in the Upper Miocene to Pliocene sediments was derived from the erosion of Mesozoic sandstones and mudstones exposed in emergent landmasses in the vicinity of the Pipiriki High and, during the Upper Miocene, along the Patea-Tongaporutu High. However the major source of illite and chlorite could have been to the east, where the Hauhungaroa-Kaimanawa Ranges formed from Mesozoic sediments now crop out and where faulting was most active from the Middle Miocene to Recent (Cope, 1965). Moreover, it is likely that the erosion of older or perhaps contemporary Tertiary lithologies, dominated by mudstones, supplied illite, chlorite, montmorillonite, kaolinite and mixed-layer clays to Upper Miocene and Pliocene sediments. Possible stratigraphic evidence for significant reworking is seen in an oil well log from the Pipiriki High area where the absence of Mt. Messenger Sandstone beds has been interpreted as due to erosion prior to Urenui sediment deposition (McQuillan, 1977). The crystalline nature of detrital chlorite in Upper Miocene to Pliocene mudstones suggests that soil profile development over this period was slight, perhaps a reflection of increasing tectonism (cf. Nelson and Hume, 1977).

The common occurrence of volcanic rock fragments throughout Upper Miocene to Pliocene sediments probably reflects erosion of older volcaniclastic sequences (e.g. Mohakatino sediments). However, tuffaceous material in Mt. Messenger (Hay, 1967) and Urenui sediments (e.g. Pl. 1.21) was probably derived largely from active volcanic centres to the west. Urenui tuffaceous sandstones resemble mineralogically the Mohakatino volcanic sandstones (cf. Table 3.6). The ubiquitous crystalline trioctahedral montmorillonite in the clay fraction of Urenui tuffaceous sandstones (Fig. 3.5; Pls. 3.26 and 3.27) is considered to have a similar origin to that in the Mohakatino Volcanic Sandstones (i.e. primarily from the alteration of volcanic material in the marine environment). The mode of occurrence of authigenic analcite in the Urenui tuffaceous sandstones (Pl. 3.27) is compatible with development from volcanic products (cf. Deffeyes, 1959; Deer et al., 1967).

Upper Miocene to Pliocene clays are unlikely to have formed during burial diagenesis as the clay mineralogy of concretions is comparable with that of host mudstones (cf. pp. 93-94). In sandstones, however, clay mineral transformations have occurred. For example, in the Mt. Messenger sandstones the low abundance and poorly crystalline nature of montmorillonite and the increased content of mixed-layer clay minerals (cf. Figs. 3.5 and 3.6) are most probably due to the degradation of montmorillonite by percolating ground waters in the highly porous sandstones (cf. Pl. 3.25).

3.3.6 PLEISTOCENE (NUKUMARUAN, CASTLECLIFFIAN AND HAWERA SEDIMENTS)

The clay mineralogy of Pleistocene sediments is dominated by illite and montmorillonite with lesser mixed-layer clays, chlorite and kaolinite (Fig. 3.5; Pls. 3.33 and 3.34). Compared with Pliocene clay assemblages, there is a marked increase in montmorillonite abundance, a decrease in the crystallinity of illite and chlorite in both Castlecliffian and Hawera sediments and an absence of chlorite from the majority of younger sediments (Figs. 3.5; 3.6 and 3.7).

Both illite and chlorite are dominantly anhedral (Pl. 3.34), suggesting they are detrital; scanning electron micrographs and thin section studies show no evidence for the authigenic development of these clay minerals. Treatment of samples with KOH (Table 2.6) suggest it is unlikely that the montmorillonite component formed by the transformation of micaceous materials. Increasing sedimentation rates in the Pleistocene (Nelson and Hume, 1977) reflect increasing tectonism at a time when depositional areas were becoming localised by basin downwarping and emerging ranges (Fleming, 1953; 1962). It is probable that illitic and chloritic clays were derived from Mesozoic sediments that formed the rising Hauhungaroa-Kaimanawa-Ruahine Ranges in the east. Subaerial erosion of Tertiary sediments emerging to the north (Nelson and Hume, 1977) could have supplied additional illite, chlorite, montmorillonite, kaolinite and mixed-layer clays to the Pleistocene sedimentary basins. Illite and particularly chlorite decrease in both abundance and crystallinity in Nukumaruan through to Hawera sediments, reflecting a relative decrease in the importance of Mesozoic and Tertiary clay mineral provenances. The decrease in crystallinity of both illite and chlorite and the virtual absence of chlorite in Castlecliffian sediments (Appendix Table IV.1) are evident in both sandstones and mudstones; hence degradational transformation of illite and chlorite in the more porous sandstone lithologies during diagenesis is not regarded a likely cause (cf. pp.91-93). It is possible that the poorly crystalline illite and chlorite represent pedogenic clays derived from soils developed on Mesozoic and Tertiary sediments. Initially this appears incompatible with the colder glacial climates of the Pleistocene (Fleming, 1962; Devereux, 1967; Hornibrook, 1970; Savin, 1977) at which time little soil development and few degradational transformations of clay minerals might be expected (cf. Millot, 1970, p.133). However the Pleistocene climate was characterised by oscillations between glacial and interglacial conditions and it is possible that during the warmer interglacial periods illitic and chloritic clays were degraded in soils developed on Mesozoic and Tertiary sediments. Post-Nukumaruan interglacial conditions characterised by summers that were warmer and somewhat drier than the present day, are indicated by the widespread, often deep, red weathering of old soils and land surfaces in the Wellington district (Te Punga, 1964; Vella et al., 1976). Fieldes (1968) showed that with weathering micas and chlorites in New Zealand Mesozoic and Tertiary sediments are increasingly altered to their less crystalline forms of other minerals, including montmorillonite. It is therefore possible that the lowered crystallinity and abundance of illite and chlorite in Pleistocene sediments reflect changes in climatic regime.

Pliocene through Pleistocene sediments contain increasing quantities of volcanogenic material, including volcanic rock fragments and plagioclase, glass shards and pumiceous debris (Appendix Table I.2, Fig. 3.4A). It is likely that the parallel increase in montmorillonite reflects a volcanic origin for some of the montmorillonite as volcanism began as early as the Upper Miocene and increased in intensity through the Pleistocene, with voluminous outpourings of acid pyroclastics in the Central North Island region (Vucetich and Pullar, 1969; Healy, 1971) and more localised intermediate eruptives in the Taranaki region (Grant-Taylor, 1964; Neall, 1972, 1974). The occurrence of α-cristobalite in the clay size fraction of Pleistocene sediments (Pl. 3.34) supports this suggestion (cf. Pl. 3.21).

3.4 STRATIGRAPHIC DISTRIBUTION OF DETRITAL, TRANSFORMED NEOFORMED CLAY MINERAL SPECIES IN THE CENTRAL WESTERN NORTH ISLAND SEDIMENTS

From the preceding sections, the relative importance of detrital, transformed and neoformed clay minerals can be assessed for a composite stratigraphic column in the central Western North Island (Fig. 3.11). The figure is presented as a visual summary for the reader and should only be used in conjunction with the foregoing and accompanying text. The clay mineralogic trends in the figure are not considered quantitative and it is recognised that any construction such as this is to some degree subjective. The relative abundance and crystallinity of the major clay mineral species in the clay fraction of the sedimentary units represent averages for that particular unit "weighted" with respect to the relative abundance of sandstones, mudstones and limestones in the unit. Thus Figure 3.11 largely represents the clay mineral trends in mudstones as they are, in general, the most abundant lithology in the study area and also contain, by volume, the most clay minerals.

Examination of the data shows that detrital, transformed and neoformed clays are present throughout the entire stratigraphic record. Detrital clays may include any of the major clay mineral species and these may exhibit varying states of crystallinity. In sediments where detrital clays are most abundant, montmorillonite and crystalline illite and chlorite are the dominant clay mineral species. Transformed clays are mostly kaolinite, illite, chlorite and mixed-layer minerals. They are most abundant in Mesozoic

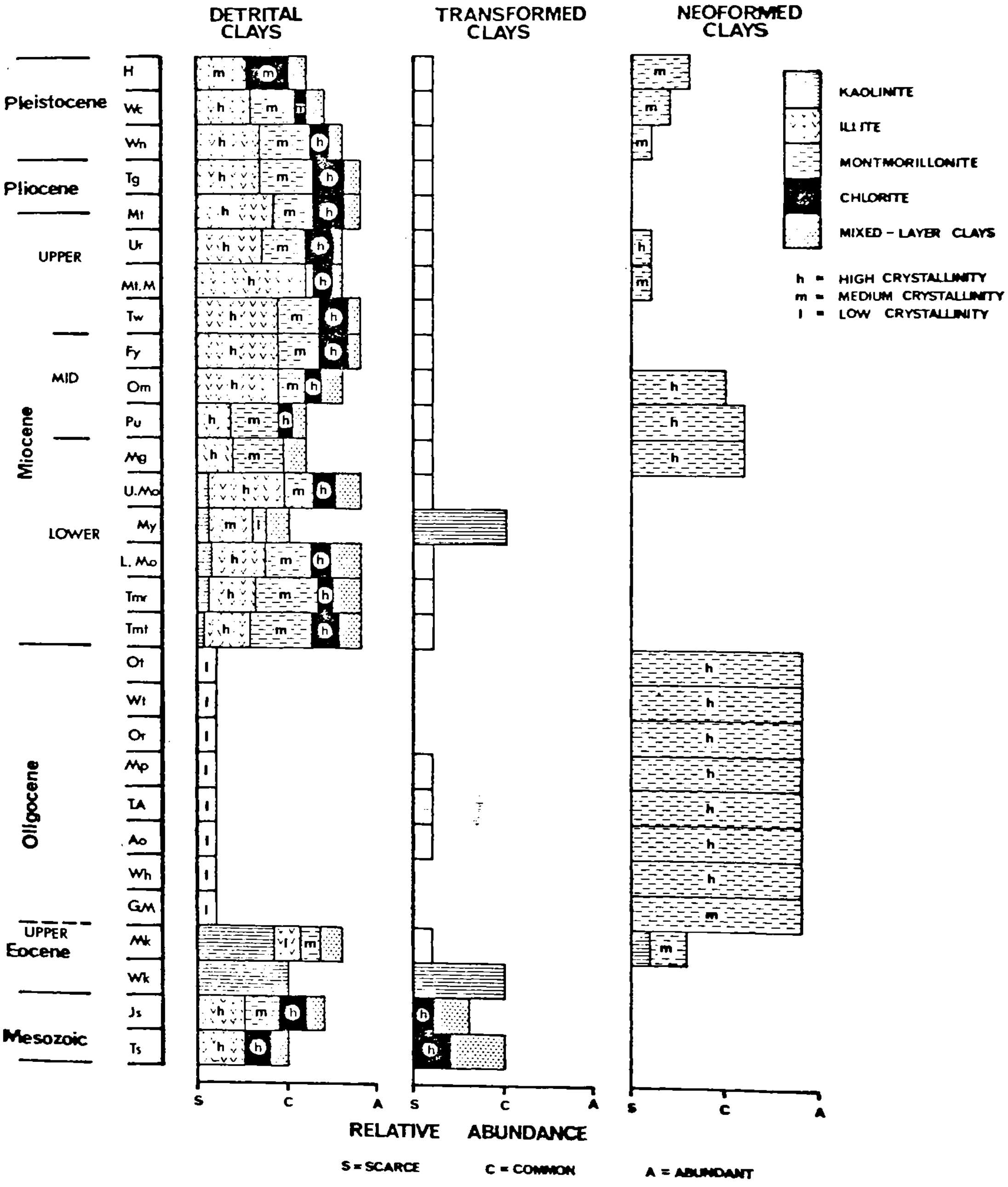


Fig. 3.11. Schematic summary of the relative importance of detrital, transformed and neoformed clays in the <2µ fraction of sedimentary units of the onshore study area. Different shading of bars denotes the major clay mineral species, and species that cannot be differentiated at the scale of the diagram are represented by unshaded bars. An overall estimate of crystallinity is given for this latter fraction where applicable.

sediments where mixed-layer clays and crystalline illite and chlorite are formed during deep burial diagenesis. Transformed kaolinite is locally abundant in Waikato Coal Measures and Maryville Coal Measures sediments. Throughout the Cenozoic stratigraphic record transformed kaolinite, illite and mixed-layer clays occur locally in porous and permeable sandstones where they have formed through post-depositional degradational alteration of existing clay minerals. Neoformed clays include rare kaolinite and abundant montmorillonite. The latter occurs at several horizons in the stratigraphic column, notably in the Oligocene carbonate-rich sediments and in volcanogenic deposits.

More detailed microscopic analysis of central Western North Island sediments, particularly of the sandstones, may establish that transformed and neoformed clays are perhaps more abundant in the column (cf. Wilson and Pittman, 1977). It is unlikely however, that they would prove to be of major volumetric importance.

3.5 FACTORS CONTROLLING THE FREQUENCY DISTRIBUTION OF CLAY MINERALS IN CENTRAL WESTERN NORTH ISLAND SEDIMENTS

The roles that provenance, climate, tectonism, the depositional environment and diagenesis have played in controlling the frequency distribution of clay minerals in central Western North Island sediments is discussed in this section and summarised schematically in Figure 3.12.

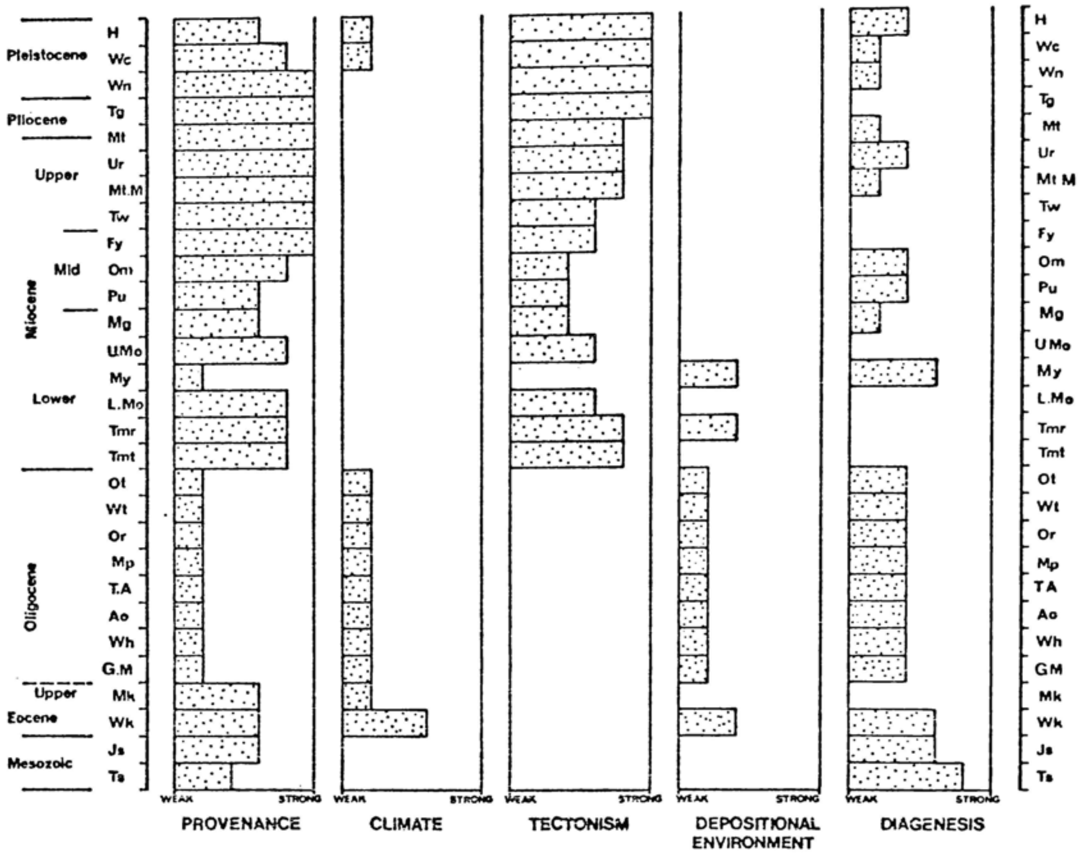


Fig. 3.12. Schematic summaries of the relative importance of the factors controlling clay mineral distributions at various stratigraphic levels in central Western North Island sediments. The graphs are constructed on the basis of data presented in section 3.5.

3.5.1 PROVENANCE

A fundamental control on the clay mineral composition of a sediment is the nature of available source material. Clay minerals may be inherited through either direct mechanical erosion of continental rocks or by reworking of soils and weathered products developed on these parent materials. Clays so derived are detrital, and the great majority of clay minerals in sedimentary rocks are probably of this origin (Weaver, 1958 a; Grim, 1968; Millot, 1970, p.33). In general provenance control is invoked when the clay mineral suite of a sediment is similar to that in "known" source rocks or sediments. In studies of Holocene sediments this approach is fairly straight-forward as the provenance is known and its contribution may be readily assessed through mineralogic comparison. In studies of pre-Holocene sediments the parent materials may still be identifiable but the weathered products and soils have usually been removed from the stratigraphic column by reworking. Thus for older sediments the situation is more complex and it is more difficult to determine whether the clays in the sediments represent those formed in the provenance areas (i.e. detrital clays) or those formed in the sediment (i.e. transformed and neoformed clays).

The nature of clay material being released from a source area is determined by the interplay of a complex set of factors including parent material composition, climate and tectonism (Folk, 1968; Millot, 1970; Pettijohn, 1975). In general terms, a more humid climate promotes soil formation and chemical weathering; clays detrital from such areas reflect the soil clay mineralogy. Conversely, increasing tectonism produces physical erosion and clays inherited by adjacent sedimentary basins tend to reflect more the source rock mineralogy. The effects of climate and tectonism in determining clay mineral composition are discussed in more detail in the following sections.

The provenance for early Tertiary sediments was initially Mesozoic sediments, but for successively younger deposits it is probable that considerable quantities of previously deposited Cenozoic clays were reworked into adjacent sedimentary basins. At certain horizons an extra-basinal volcanic provenance has been important in providing material to nearby depocentres and surrounding areas, where glassy constituents were subsequently altered to montmorillonite.

Mesozoic sediments are characterised by crystalline illite and chlorite and by mixed-layer clays in the clay fraction. The dominance of montmorillonite in the clay fraction of overlying Oligocene sediments indicates these clays were not directly inherited from Mesozoic sediments. Furthermore it is considered unlikely that the majority of montmorillonitic clay material originated through weathering and pedogenesis (cf. pp.103-111). Taumatamaire (early Lower Miocene) sediments are rich in montmorillonite derived largely from Oligocene sediments. By comparison larger quantities of more crystalline illite and chlorite in Taumarunui (early Lower Miocene) sediments reflect an increasing contribution from a Mesozoic provenance to the east and north of the basin. Post-Mahoenui Group (late Lower Miocene) sediments contain increasing quantities of crystalline illite and chlorite reflecting a parallel increase in the Mesozoic sandstone and mudstone provenence with decreasing age. This general trend is complicated by local diagenetic effects, volcanism and reworking of previously deposited Tertiary sediments.

The degree of induration in the source area may have been important in accentuating provenance control in the post Oligocene clay mineral suites. Field observations indicate that Cenozoic mudstones in the study area are generally highly susceptible to erosion (e.g. Pls.1.4, 1.18 and 1.23) and mineralogic analyses suggest that the erosion products from these sediments represent the parent materials rather than the soils (cf. pp.230-231). The high erosion potential of Cenozoic mudstones may have been a significant factor contributing to their reworking into younger sediments.

3.5.2 CLIMATE

Climate determines the clay mineralogy of sediments by controlling the development of clays during weathering and soil formation and thus the nature of material being released from a source area. Rateev et al.(1969) demonstrated that the distribution of clay minerals in the world's ocean sediments is closely related to geographical latitude and hence climate. Numerous works have emphasised the importance of climate in soil forming processes. Climate is one important factor in determining the leaching environment of soils and thus their clay mineralogy, although the time over which the climatic regime acts on a given parent material is important in determining the final product (Barshad, 1964, p.41-44 and 1966; Barnhisel and Rich, 1967; Birkeland, 1969). Under an arid climate, whether hot or cold, chemical weathering is typically minimal because of the lack of water; hydrolysis is reduced and soil formation is slight. Physical weathering predominates and clay minerals released by erosion reflect source rock composition. On the other hand, a humid climate gives rise to deep soils because of intensive hydrolysis and the presence of abundant vegetation which serves to 'anchor' the soil profile as well as catalyse certain chemical and biochemical reactions (Millot, 1970, pp.109-134). Alteration of parent material is accomplished through mineral transformations and by neoformation of clay material. Between the arid and humid climate end-members there is a spectrum of possible products whose composition are dependent on parent material, climatic and time factors. If for instance orogenic forces maintain geologic uplift and erosion at a high rate, then no matter how conducive the climate is to soil formation, extensive soil cover is unlikely to be established or retained.

In central Western North Island sediments, climatic control on clay mineralogy can be demonstrated at only three stratigraphic levels (cf. Fig. 3.12). Elsewhere the climatic effects are uncertain mainly due to the masking of climatically determined clays by those controlled by other factors. Waikato Coal Measures mineralogy is characterised by stable minerals such as quartz and kaolinite and rare degraded minerals such as poorly crystalline chlorite. The sediments had as their only source material Mesozoic sediments. Fieldes (1968) showed that intense weathering and leaching of Mesozoic sediments under present day conditions results in a kaolinitic clay mineral suite. This suggests that the extensive kaolinitic clays of the Waikato Coal Measures may have formed at a time when such conditions were widespread and that climate was perhaps important in determining the clay mineralogy. However kaolinite is primarily an indicator of an acid leaching environment where it forms by authigenesis (Millot, 1970, p.164). The abundant carbonaceous material in the Waikato Coal Measures would have provided an environment conducive to kaolinite authigenesis. Thus clay mineral data alone do not demonstrate a climatic control. Other evidence, however, suggests that climate played an important role in forming Waikato Coal Measures clays. The Cretaceous to early Tertiary kaolinite-rich regolith developed on Mesozoic sandstones and mudstones in many parts of New Zealand is indicative of a climatic control for kaolinite formation. The climate through this period probably was warm temperate to marginally tropical (cf. pp. 99-100) and would have provided conditions conducive to kaolinite development (cf. Rateev et al. 1969). Moreover the long period of time (up to 90 m y) over which these conditions prevailed would have been sympathetic to kaolinite formation (cf. Birkeland, 1969).

In Oligocene times, the small total volume of terrigenous material indicates regional stability on the adjacent landmass (Nelson, 1973) and the highly calcareous nature of the sediments testifies to a dominantly chemical sedimentary regime. These features, and the bulk and clay mineralogy of Oligocene sediments suggest that weathering and erosion of Mesozoic sediments on the adjacent landmass produced primarily dissolved materials and highly degraded clays (cf. pp. 99-100). The warm temperate climate suggested for the Oligocene would only partially favour such reactions which are most intense under a humid tropical climate

(Millot, 1970, p.99-134). For this reason tectonic quiescence may have been an important factor in determining the nature and supply of materials to the depocentres by allowing time for pedogenic dissolutions to take place (cf. Birkeland, 1969).

Pleistocene sediments show a marked decrease in illite and chlorite crystallinity with decreasing age. There is little evidence to suggest that this change results from diagenetic effects or from a change in source area (cf. p.136). Decreasing crystallinity of illite and chlorite commonly parallels an increasing intensity of soil forming processes (Fieldes, 1968; Millot, 1970). For Mesozoic parent materials, these changes could be accomplished by a change to a warmer climate (Fieldes, 1968), perhaps associated with the Pleistocene interglacial periods (cf. p.136).

With decreasing age, Cenozoic sediments of the central Western North Island show a marked change from a predominantly authigenic mineral assemblage characterised by crystalline montmorillonite to a predominantly detrital mineral assemblages characterised by crystalline illite and chlorite (Figs. 3.5 and 3.6; Section 3.3). Similar patterns in clay mineral distributions observed in deep-sea sediments have been used by a number of workers to demonstrate broad trends in climatic change. Heath (1969) described deep-sea sediments of Eocene to Holocene age in the Equatorial Pacific and noted that the middle Eocene (50 m.y.) to late Miocene (10 m.y.) assemblage of montmorillonite - intermediate plagioclase - zeolites in the central Pacific was replaced in younger sediments of the more northern Pacific by a quartz-crystalline illite assemblage. He considered the latter assemblage was derived from the subaerial erosion of rockflour produced as a result of climatic changes associated with continental glaciation. In the North American Basin of the Atlantic Ocean the mineralogic difference between the montmorillonite-zeolite Cretaceous sediments and the disconformably overlying chlorite-illite Quaternary sediments have been interpreted as due to climatic factors (Jacobs, 1970). Jacobs and Hayes (1972) demonstrated that sediments of the Equatorial Pacific from 11 to 3 m.y. B.P. have a montmorillonite-rich clay mineral assemblage (montmorillonite 80%, illite 10-15%, and kaolinite plus chlorite 5%), while in the sediments younger than 3 m.y. B.P. the clay mineral assemblages are more detrital, with increasing quantities of crystalline illite and chlorite. They suggested this change was the result of dilution of montmorillonite by an influx of detrital material from increased physical weathering, surface run-off, and wind activity associated with an intensification of continental glaciation. More recently Jacobs (1974) has described the mud fraction mineralogy of Eocene to Holocene Antarctic deep-sea sediments. The Eocene to Oligocene clay fraction is dominated by montmorillonite (montmorillonite 80%, poorly crystalline illite 15%, and chlorite plus kaolinite 5%) while the silt fraction contains an authigenic mineral assemblage with zeolites, barite and calcite. During the Miocene there is a decrease in montmorillonite, an increase in the abundance and crystallinity of illite, an increase in the content of chlorite and kaolinite, and the silt fraction becomes largely detrital in nature, containing quartz, plagioclase, microcline, chlorite, muscovite and amphibols. During the late Miocene (7-10 m.y.) a consistent abundance pattern emerges which has remained constant for the last 5 m.y. of geologic time. These changes were interpreted by Jacobs (1974) as due to the dilution of Eocene and Oligocene authigenic mineral assemblages by increasing quantities of detrital minerals produced by intense physical erosion of the Antarctic continent by the rapid build-up of massive ice sheets prior to the late Miocene. Gostin and Moriarty (1975) describe the clay mineralogy of two cores from south of Tasmania and close to the Antarctic continent (D.S.D.P. Leg 29, Site 280) in which there is a change from montmorillonite to predominantly illite-chlorite clay mineral assemblages in the late Eocene to early Oligocene, suggesting a change from temperate source area conditions to cold glacial conditions at that time.

Thus the clay mineral distributions in deep-sea sediments commonly show a change from predominantly authigenic to predominantly detrital clay minerals as a result of world-wide climatic deterioration associated with the Cenozoic build-up of the Antarctic ice-sheet. The timing of this event varies mainly from the Middle and Upper Miocene in the Antarctic area (Jacobs, 1974), to the Pliocene in the Equatorial Pacific region (Heath, 1969; Jacobs and Hayes, 1972), but occurred as early as the late Eocene in sites which at that time lay close to the Antarctic continent (Gostin and Moriarty, 1975). Any direct comparison of the deep-sea sediment data with that obtained from the Cenozoic sediments in this study must be made with caution since variations in provenance can affect clay mineral distributions in the sediments, even over short distances (cf. pp.191-240). However, one striking similarity is noted with the Antarctic deep-sea sediment data and that is the replacement of an Oligocene authigenic mineral assemblage by predominantly detrital assemblages in post-Oligocene times (Appendix Table III.1 and IV.1; Figs.3.4, 3.5 and 3.6). It is tempting in the first instance to attribute this marked change to climatic deterioration, as was suggested for deep-sea sediments. On closer examination, however, this explanation is unlikely because:

(1) the influx of detrital material occurs in the Lower Miocene which is earlier than that in the Antarctic deep-sea sediments which lay, moreover, at higher and therefore relatively colder latitudes;

(2) paleotemperature data from New Zealand Tertiary sediments (Devereux, 1967; Hornibrook, 1971) indicate that appreciable cooling did not take place until the Upper Miocene (i.e. 10 m.y. B.P.); and

(3) available evidence suggests a subtropical Lower Miocene climate(cf. p.123).

It is not likely that diagenetic alteration would offer an alternate explanation for the changes in clay mineral abundances observed in Cenozoic sediments. Mudstones show little evidence for diagenetic changes in clay mineralogy and in fact their clay mineral suites closely approximate those existing at, or soon after, the time of sediment deposition (cf. pp.93-94).

In summary, therefore, climate was an important factor in determining clay mineral assemblages at certain periods in central Western North Island sediments, especially in the Upper Eocene Waikato Coal Measures and Pleistocene sediments, but overall it does not appear to have been an important control in determining bulk clay mineralogic characteristics.

3.5.3 TECTONICS

It is fairly well established that while many factors affect sedimentation, and hence the bulk characteristics of sedimentary rocks, the most fundamental is probably tectonics (Krumbein and Sloss, 1963; Pettijohn, 1975). The nature of a sediment is determined not only by the intensity of formative processes acting on it but also by the time or duration that the processes are active (Pettijohn, 1975, p.571-584). Tectonism controls the residence time of sediments in the source area, and thus the degree to which they are weathered, the rate of subsidence in basins of deposition and, to a lesser degree, the diagenetic modification of sediments through folding and faulting. Moreover, while the textural maturity and mineralogy of a sediment are dependent mainly on environment and source area lithology respectively, the volumetric importance of specific sedimentary environments and source rocks is profoundly affected by tectonism (Folk, 1968). When tectonics are active, slopes are generally too steep and the energy of waters too vigorous to permit significant widespread development of soils. Under these conditions mechanical erosion is predominant and weathering is slight. Erosion products are rapidly transported to depositional basins where sedimentation is overwhelmingly detrital and mineral suites are characterised by both stable and unstable mineral types and rock fragments. Clays are typically more crystalline and their composition reflects provenance. A variant of the above process is tectonic rejuvenation that initiates erosion after a long period of tectonic calm. Deep soils developed during the quiescent period are eroded and swept to the sedimentary basins and sedimentation is again detrital. When tectonics are slight soil formative processes have much longer to operate. Under temperate and tropical climates luxuriant vegetation develops protecting the soil profile, dissolution is active and the dissolved products and degraded minerals are transported to sedimentary basins where sedimentation is predominantly chemical. The resulting mineral suites are characterised by a few stable minerals, such as quartz, and by chemical products such as calcium carbonate, α-cristobalite, neoformed montmorillonite and glauconite. Between these end members moderate tectonics and weathering yield a spectrum of mineral products including transformed clays, whose development is governed by the relative intensities of tectonism and climate in the source area.

The effects of tectonism in controlling the clay mineralogy of Mesozoic sediments are not discussed because too few Mesozoic samples were analysed and because it was earlier shown that their clay mineralogy has been significantly affected by burial diagenesis (cf. pp. 94-97).

Existing evidence concerning the importance of tectonism in controlling sedimentary patterns at certain stratigraphic levels in the Tertiary of the study area is summarised below. Kear and Schofield (1959) and Nelson (1973) showed that the Oligocene was characterised by dominantly chemical-biochemical sedimentation, low rates of terrigenous sedimentation and by-passing and erosion of sedimentation on the sea floor. Nelson (1973) equated the small total volume of terrigenous material in the Oligocene Te Kuiti Group with regional stability in the South Auckland area at that time. In Miocene to Pliocene sediments the relationship between fault strike modes and the geologic age of affected strata suggests that the location of depocentres was governed by differential movements of sub-basin fault-blocks during deposition (Hay, 1967; Cope and Reed, 1967; Nelson and Hume, 1977), coupled with a general southerly shift in the zone of maximum sediment accumulation with time (Hay, 1967). During Mahoenui times (early Lower Miocene) the very thick sequences (up to 1384 m) of sediments at and east of Ohura resulted from a substantial downfaulting along the eastern side of the Ohura Fault (Fig. 3.3; Glennie, 1959; McQuillan, 1977), while the coarser textured and more detrital nature of the most western Taumatamaire sediments reflects continued movement of the Herangi High (Happy, 1971; Nelson, 1977; pp.115-116). In the Mokau times (late Lower Miocene) the restriction of sediments to the area west of the Ohura Fault, the rapid thickness changes towards the fault, and the increased incidence of conglomerates in the same direction all provide evidence of a control on Mokau sedimentation by movements on the Ohura Fault (Hay, 1967). The removal of some 300 m of Mahoenui beds prior to Mokau deposition in the area east of Ohura (Stainton and Gibson, 1964) supports considerable uplift and reworking. In Middle to Upper Miocene sediments, locally developed but spectacular intraformational subaqueous slump structures and conglomerate-filled channels have been attributed to periodic movements of the presently subsurface Patea-Tongaporutu Basement High (Glennie, 1958; Hay, 1967). The general southerly shift in the zone of maximum sediment accumulation with time, as evidenced by sediments younging to the south (Fig. 1.3), culminated in rapid downsinking and sedimentation in the South Wanganui Basin where over 4 km of sediments accumulated in Pleistocene times (Fleming, 1953).

Using field and laboratory data from this study, together with some additional data from Henderson and Ongley (1923), Hay (1967), Katz (1968), Haddock (1970), Happy (1971) and Nelson (1973), sedimentary features characteristic of contrasting tectonic regimes are compared using sedimentary logs for Cenozoic sediments of the study area to gauge the relative importance of tectonism in controlling the sediment properties.

Sedimentary features indicative of tectonically active versus tectonically quiescent environments have been described in Folk (1968), Millot (1970) Blatt et al. (1972), Pettijohn (1975) and several features related to this work are summarised in Table 3.7. It is recognised that the Cenozoic sediments are commonly diachronous, show

	Tectonically Unstable	Tectonically Stable
1.	High relief	Low relief
2.	Thick sedimentary units	Thin sedimentary units
3.	High sedimentation rates	Low sedimentation rates
4.	Many mineral types, both stable and unstable; detrital materials e.g. quartz, feldspar and/or rock fragments, crystalline illite and chlorite	Few mineral types, only stable ones and chemical products e.g. quartz, secondary clays (glauconite, neoformed crystalline montmorillonite), $CaCO_3$, α-cristobalite
5.	Litharenites-lutites-muddy rudites	Quartz arenites-limestones-coals and associated kaolinitic fireclays
6.	Immature sediments; muddy matrix; unsorted sediments	Mature sediments; pore spaces; cement; sorted sediments
7.	Fossils rare	Fossils common
8.	Slumping, redeposition, turbidity currents	Wave action, tidal currents
9.	Graded bedding	Cross-bedding, horizontal stratification
10.	Volcanic activity common	Volcanic activity rare or absent

Table 3.7. Sedimentologic features characteristic of tectonically stable and unstable sedimentary basins (after Folk, 1968; Millot, 1970; Blatt et al., 1972; Pettijohn, 1975).

rapid lateral facies changes and are internally varied. However, for this comparison the data are purposely generalised to highlight only the major, more regional sedimentary characteristics of the formations in the sequence. It is emphasised that the sedimentary features in Table 3.7 must be considered collectively in order to evaluate the relative importance of tectonics during sedimentation.

3.5.3A SEDIMENTARY LOGS

The bulk mineral content of the sedimentary units varies with lithology (cf. Figs. 3.4A, 3.4B, 3.5 and 3.6). To compensate for this, and to obtain "truer" formational averages, percentages plotted on the compositional logs are weighted according to bulk lithologic content of individual formations (Appendix Table IX.1).

A short discussion of each of the sedimentary logs follows.

SEDIMENT LITHOLOGY

The Tertiary formations consist of varying proportions of mudstones, sandstones and limestones, with rare coal beds (Appendix Table I.2; Figs. 3.13 and 3.14). Apart from the

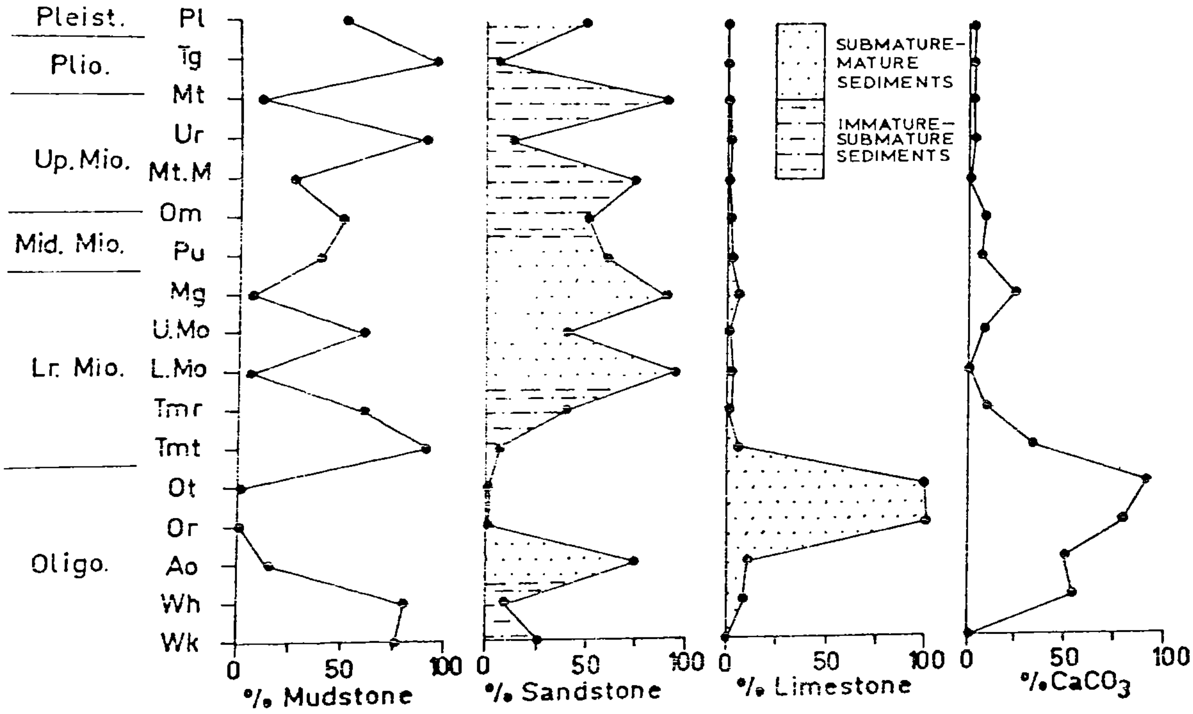


Fig. 3.13. Variations in sediment lithology and textural maturity in Cenozoic sediments of the study area.

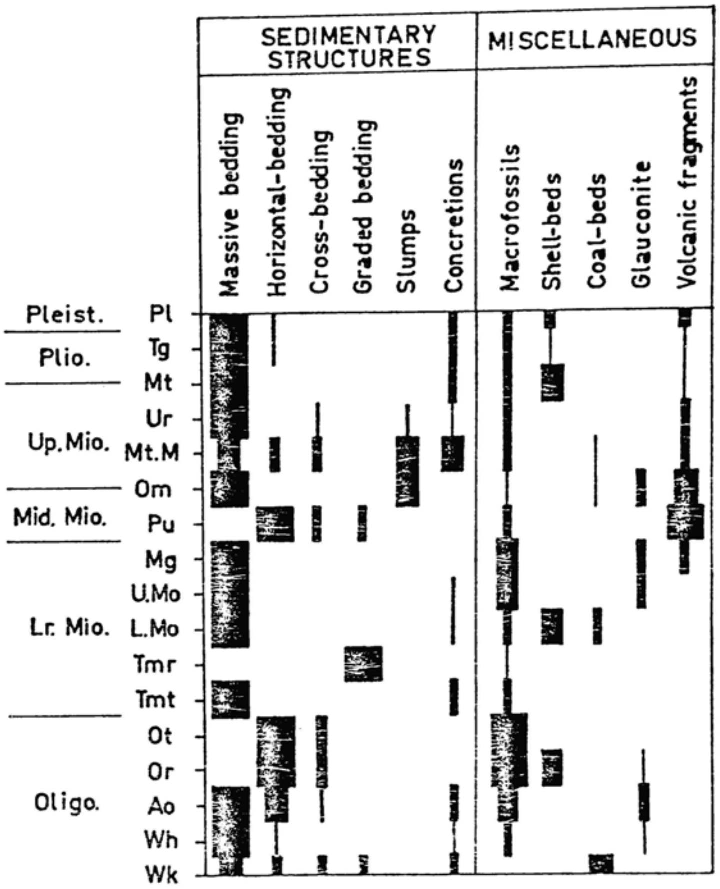


Fig. 3.14. Bulk variations in selected sedimentologic properties in Cenozoic sediments of the study area.

basal coal measures (Pl. 1.2) the Oligocene formations are typified by highly calcareous mudstones and sandstones (Pls. 1.3 and 1.4) and by the widespread development of coarse skeletal limestones (Pls. 1.6 and 1.8). Miocene to Pleistocene sediments are, in contrast, predominantly non-calcareous, with rare thin limestones, and molluscan shell-beds and calcareous concretions (Figs. 3.13 and 3.14). Although mudstone to sandstone abundance varies erratically in Miocene to Pleistocene sediments the overall relative abundance of the two lithologies is similar. In the Cenozoic sequence sandstones of Middle to Upper Oligocene and late Lower to Middle Miocene are notably the most texturally mature (Fig. 3.13).

Cenozoic sediments are commonly massive (Fig. 3.14; Pls. 1.12, 1.13 and 1.18 - 1.24) owing to the deposition of predominantly fine terrigenous material under uniformly low energy conditions and homogenisation of bottom sediments by burrowing organisms (cf. Byers, 1974). Horizontal-and cross-bedding, indicative of significant tractive current action, are common in Oligocene and Middle Miocene formations but generally rare in other sedimentary units (Fig. 3.14). Graded bedding and flysch-type sequences (Bouma, 1962) typify Taumarunui Formation sediments (Fig. 3.14; Pl. 1.10), while the occasional graded beds in Waikato Coal Measures and Purupuru volcanic sandstones reflect sorting by river currents (Nelson, 1973) and shallow marine bottom currents respectively.

Flysch-like sequences (Van der Lingen, 1968) occur infrequently through the late Lower Miocene and Upper Miocene to Pleistocene formations. Intraformational slump structures are most common in the Omoao Formation and Mt. Messenger Sandstone.

Macrofossils, including coarse shell hash, are common to abundant in the Oligocene and late Lower Miocene formations but scarce elsewhere (Fig. 3.14). The distribution of glauconite pellets (Fig. 3.14) roughly parallels that of shell fragments. Volcanic rock fragments are most abundant in Middle Miocene volcanogenic sediments (Fig. 3.14).

THICKNESS AND SEDIMENTATION RATE

Pettijohn (1975, pp.571-572) suggested that tectonism was the most fundamental control on sedimentation mainly through its influence on total sediment supply and rate of subsidence so that variations in the values for thickness and sedimentation rate should closely approximate variations in the degree of tectonic activity.

From formation thicknesses (Tables 3.2 - 3.4) and age (Appendix Table IX.1) data "rock" sedimentation rates have been estimated for the sedimentary units on the basis of absolute ages proposed for the New Zealand Cenozoic stages by Berggren (1972). It is recognised that some limits are imposed on the accuracy of these data as:

1. Deposition was not continuous for the full time span of formations as there were periods of non-deposition and/or erosion.

2. Some formations are partial time equivalents of others e.g. Aotea Sandstone and Orahiri Limestone; Taumatamaire and Taumarunui Formations.

3. Pre-lithification sedimentation rates might be up to a maximum of 50% greater than those values quoted for "rock" sedimentation rates (Nelson, 1973, p.556).

For these reasons the data are in no way construed as being absolute but are probably valid for displaying major trends in sedimentary rates with time.

Each of the Oligocene and late Lower Miocene to Middle Miocene formations are typically <50 m thick, while each of the early Lower Miocene and Upper Miocene to Pleistocene formations have thicknesses of a few to several hundred metres (Appendix IX.1; Fig. 3.15). Calculated

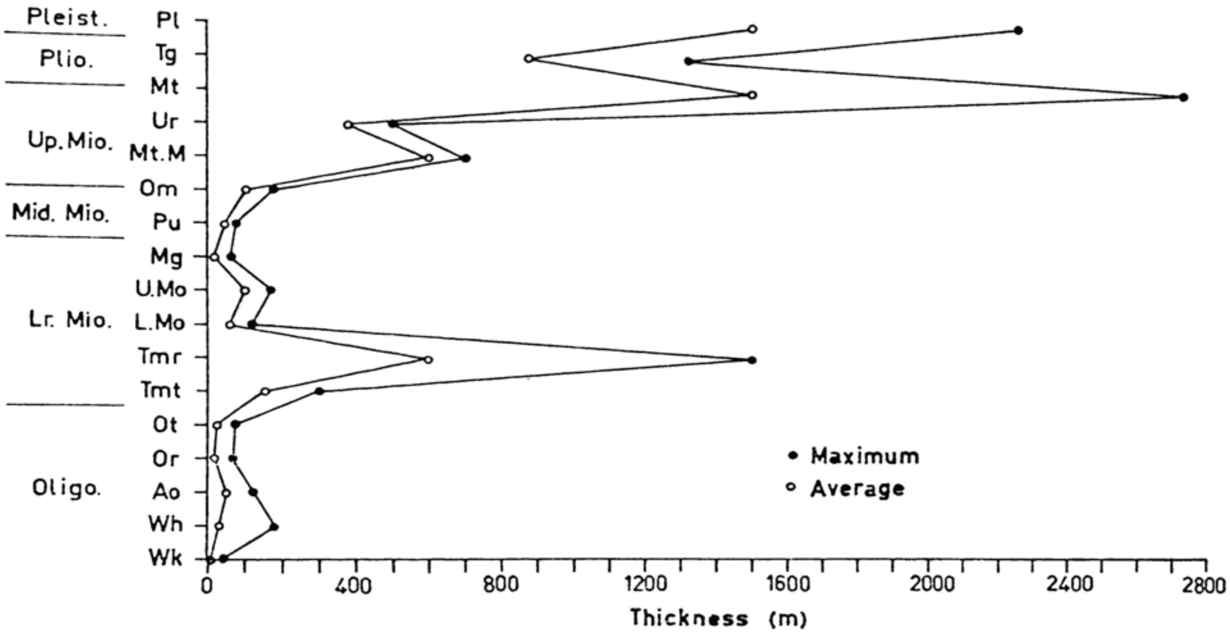


Fig. 3.15. Variations in sediment thickness for Cenozoic sediments of the study area.

"rock" sedimentation rates parallel the thickness curves, with high to very sedimentation rates in the early Lower Miocene and Upper Miocene to Pleistocene, and low to very low rates in the Oligocene and late Lower Miocene to Middle Miocene (Appendix IX. 1; Fig. 3.16).

BULK MINERALOGY

Quartz abundance is generally low in Oligocene and Middle Miocene sediments because of calcite and plagioclase dilution respectively (Fig. 3.17). In the acid insoluble residue of Oligocene calcareous sediments, however, quartz abundance is notably high (Fig. 3.17). Peak abundances for quartz commonly coincide with those formations dominated by sandstone lithologies.

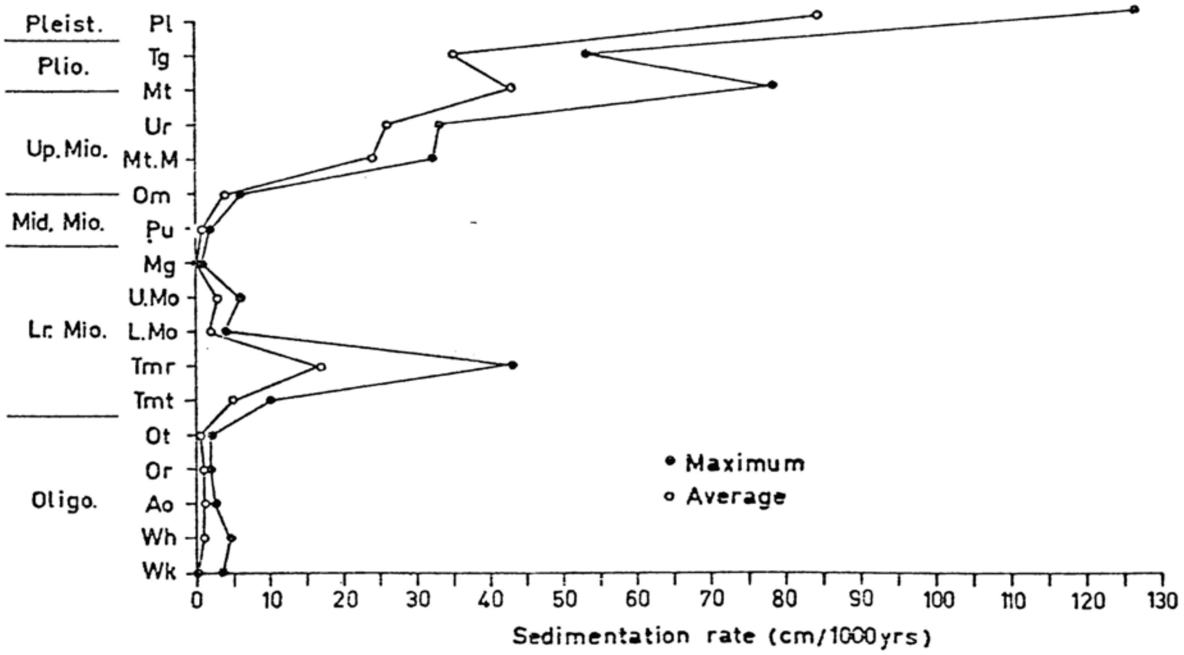


Fig. 3.16. Variations in "rock" sedimentation rates for Cenozoic sediments of the study area. Additional "lift-out" copy available in back pocket.

Plagioclase feldspar abundances in samples generally increases with decreasing age and is notably most abundant in Middle Miocene volcanogenic sandstones (Fig. 3.17). The quartz/feldspar ratio shows an overall decrease with decreasing age with values greater than one being restricted to the Oligocene formations.

Calcium carbonate is most abundant in Oligocene sediments, where it occurs primarily as skeletal fragments (Fig. 3.17). Calcium carbonate is rare to common in the Lower and Middle Miocene formations but is consistently rare in Upper Miocene to Pleistocene formations. The moderately high calcium carbonate in the early Lower Miocene formation results largely from reworking of calcareous Oligocene sediments (cf. p.115).

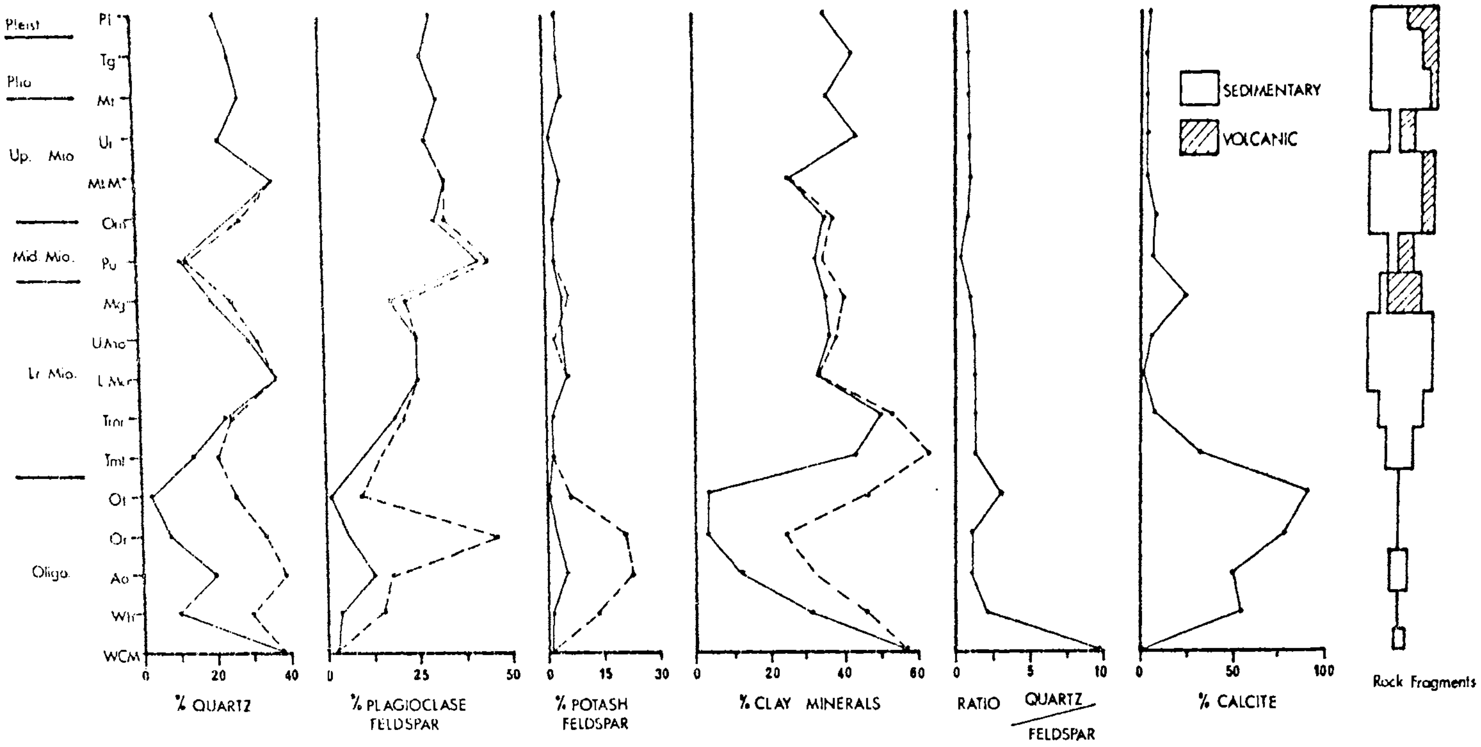


Fig. 3.17. Variations in bulk mineral abundance in the total sample (solid lines), in the acid insoluble residues (dashed lines) and in rock fragment abundance for Cenozoic sediments of the study area.

Rock fragments are most abundant in post-Oligocene sediments. The ratio of sedimentary to volcanic rock fragments is high in Oligocene and Lower Miocene sediments (Fig. 3.17).

CLAY MINERALOGY

Clay minerals are the major mineral constituents in many sedimentary units, although the total abundance of clay minerals is dependent on bulk lithology and on the rock fragment and glauconite content of samples (Figs. 3.17). Bulk clay mineral abundance shows a notable increase across the Oligocene/Miocene boundary, after which abundances remain generally similar (Figs. 3.17).

Average clay mineral abundance and crystallinity in the < 2µ size fraction of sediments from the Cenozoic sequence is summarised in Fig. 3.18.

Kaolinite is abundant in Upper Eocene sediments but rare to scarce in younger lithologies. Crystalline montmorillonite is relatively abundant in Oligocene and early Lower Miocene sediments and common in Middle Miocene sediments. The relative abundance of crystalline montmorillonite in early Lower Miocene sediments results largely from reworking of montmorillonite-rich Oligocene sediments (cf. pp.112-119). Illite and chlorite are most abundant and most crystalline in Miocene to Pleistocene sediments. The decrease in both illite and chlorite abundance in Middle Miocene sediments is attributed largely to dilution by volcanogenic crystalline montmorillonite (cf. pp.124-130). Mixed-layer clays are most abundant in middle Lower Miocene to middle Upper Miocene sediments.

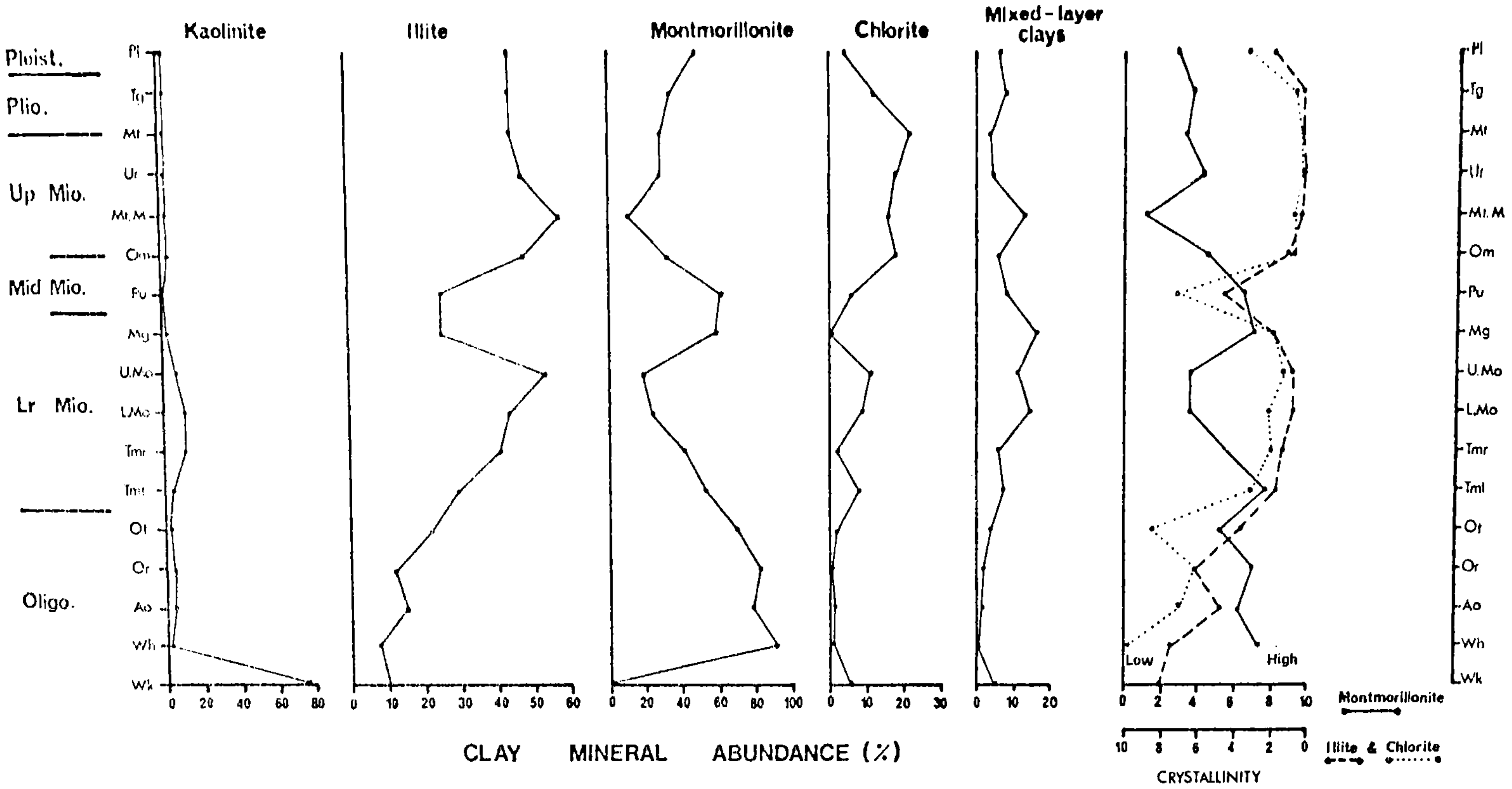


Fig. 3.18. Average clay mineral abundance and crystallinity of clay mineral species in the 2u size fraction of sediments from the major sedimentary units in the study area. Abbreviations of formation names defined in Table 1.1.

SYNTHESIS

A brief interpretation of the sedimentary logs in terms of those sedimentologic features characteristic of tectonically active and quiescent sedimentary basins (Table 3.7) follows.

Relative tectonic quiescence is suggested for Oligocene times by the highly calcareous and fossiliferous, texturally mature, and locally coaly and glauconitic Oligocene sediments (Figs. 3.13 and 3.14). The generally low carbonate and high terrigenous grain content of post-Oligocene formations is indicative of increasing tectonism over this time. Periods of more intensified movements early in the Lower Miocene and in the Upper Miocene to Pliocene are suggested by the texturally immature character of formations of these ages, and by the local occurrence of flysch-type and syndepositional slump structures in the same formations. The sediment thickness and sedimentation rate curves (Figs. 3.15 and 3.16) strongly suggest increased tectonism in the early Lower Miocene and Upper Miocene to Pleistocene periods. Low rates of tectonism in Oligocene sediments are indicated by the high carbonate content of the bulk samples and by the relatively high quartz content in the acid insoluble residues. The high quartz content, along with the presence of scarce degraded illite and chlorite in the clay fraction, results from the more complete weathering of unstable minerals in Mesozoic source rocks under tectonically quiescent conditions. Abundant crystalline montmorillonite was neoformed in marine sedimentary basins, characterised by low rates of sedimentation, from products derived by the weathering of Mesozoic materials on an adjacent tectonically stable landmass. Compared to Oligocene lithologies younger sediments have, in general, more diverse mineral assemblages including quartz, plagioclase feldspar, rock fragments and a clay mineralogy characterised by detrital crystalline illite and chlorite. The mineralogy is compatible with derivation from a Mesozoic provenance which is generally tectonically active and in which mechanical erosion predominates. The moderate calcite and montmorillonite contents of the early Lower Miocene formations is anomalous insofar as much of the carbonate and clays were probably derived from the reworking of highly calcareous, montmorillonite-rich Oligocene sediments (cf. pp.112-119). The peak plagioclase and montmorillonite values in Middle Miocene sediments are the result of contemporaneous volcanism.

Collectively the sedimentary log data show that the change from a predominantly authigenic Oligocene clay mineral assemblage to a predominantly detrital Miocene to Pleistocene clay mineral assemblage is paralled by an increase in tectonism. The data indicate the relative intensity of tectonism has not been uniform but shows an overall increase in intensity with decreasing age. Previous discussions gave reasons as to why the(se) change(s) was unlikely to have been determined by climate or diagenesis. In view of the considerable evidence that tectonism influenced other sedimentary properties it is highly likely that increasing tectonism played a major role in determining the post-Oligocene clay mineralogic character. A more detailed interpretation of the tectonic implications that can be inferred from the relationships between sedimentary properties and tectonics is given in Nelson and Hume (1977; see back pocket).

3.5.4 DEPOSITIONAL ENVIRONMENT

The relationship of clay minerals to depositional environments has been described by several authors (e.g. Grim, 1968; Keller, 1970; Millot, 1970), but the effectiveness of clay minerals as environmental indicators is still debatable. Clay mineral environmental relationships have been inferred largely from studies of Recent sediments and this information has been extrapolated to ancient sediments. Weaver (1958a, 1967), on the basis of hundreds of analyses of clays from ancient sediments showed that no single clay mineral is specific to any one environment, even in such broad terms as marine and non-marine. His data do show however that illite and montmorillonite are more characteristic of marine sediments, whereas kaolinite is more typical of continental, and especially fluvial deposits. In a review of clay mineral distributions in ancient and Recent sediments Millot (1970) supported these findings, Keller (1970) reported similar results and stressed that the use of clay minerals in paleoenvironmental interpretation was extremely limited unless the clay minerals 'formed there' could be differentiated from those 'found there'. He emphasised that clay mineral data cannot be used on their own in paleoenvironmental interpretation and that as far as possible investigators should use clay mineral data in conjunction with other techniques of environmental analysis.

The depositional or local environment exerts a control on clays through a number of physical and chemical factors (Pettijohn, 1975, pp.532-542). Physical factors such as current type (air, water, ice, etc.), current energy, direction and persistence operate to sort and redistribute material arriving in the environment. Chemical factors, including salinity, temperature and pH, modify clay minerals in the depositional basins if they are unstable under that chemical regime. In addition, biologic factors may be important, such as sediment reworking by burrowing organisms and the production of reducing microenvironments associated with the decay of organic debris.

Both previous literature (Henderson and Ongley, 1923; Hay, 1967; Katz, 1968; Haddock, 1970; Happy, 1971; Nelson 1973) and field data indicate the Cenozoic sediments of the central Western North Island are primarily shelf deposits with rare marginal marine and fresh water beds. Earlier (pp.91-93) it was shown that because post-depositional alteration of clay minerals is relatively common in porous sandstones only the clay mineral data from mudstones and limestones should be used for paleoenvironmental analysis. Fig. 3.19 illustrates schematically the types and relative amounts of clay minerals found in the major depositional environments in central Western North Island Cenozoic sediments. The clay mineral data show that:

1. No one clay mineral is restricted to a particular environment.

2. Kaolinite is most common in fluvial sediments while illite and montmorillonite and, to a lesser extent,

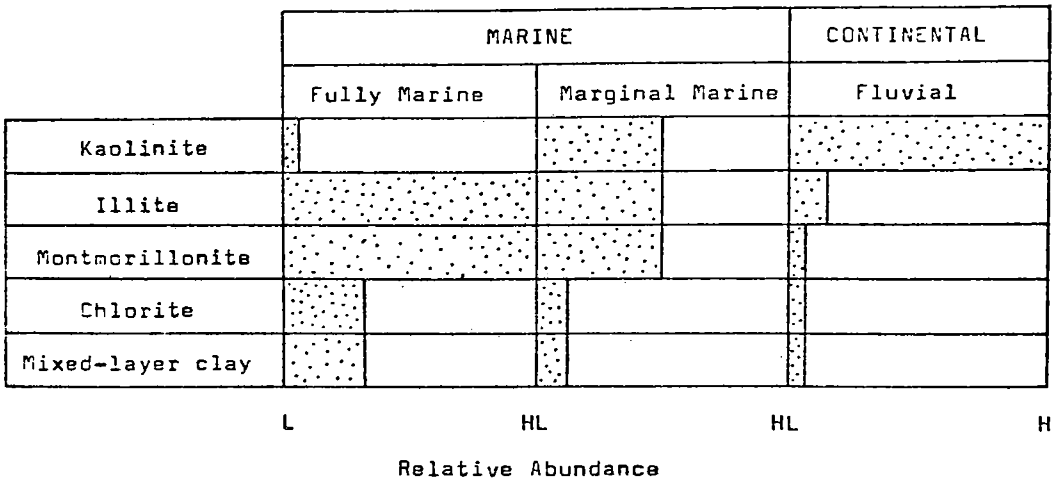


Fig. 3.19. Summary of the types and relative abundance (L = low; H = high) of clay mineral species occurring in the major depositional environments for the Cenozoic sequences in the central Western North Island. Data from Appendix Table IV.1.

chlorite and mixed-layer clays are most abundant in marine sediments.

3. Clay mineral assemblages in any one major depositional environment are, in fact, polymineralic. Further, no specific relationship exists between clay mineralogy and sediment lithology (cf. Fig. 3.5), keeping in mind, however, the possibility of the diagenetic changes that can occur particularly in sandstones.

The restriction of abundant kaolinite to 'known' fresh water deposits of the Waikato and Maryville Coal Measures suggests that, for these sediments at least, kaolinite is characteristic of a non-marine environment.

In support of this the mudstone fabrics in the Waikato Coal Measures show the clays were deposited in a non-flocculated state, presumably therefore under fresh water conditions.

As kaolinite is most abundant in continental sediments it is highly likely that it has the potential to be reworked into adjacent marginal marine sediments. Mangakotuku siltstones contain a marginal marine fauna (cf. p.101) and the generally abundant kaolinite in the clay fraction is compatible with such an environment of deposition. More detailed work by Kear and Schofield (1959) and Nelson (1973) has shown the Mangakotuku Siltstone has, in fact, many features characteristic of estuarine or lagoonal environments. Kaolinite is also common in marginal marine carbonaceous siltstones and conglomerates of the Lower and Upper Mokau Groups. The recognition of abundant kaolinite in marginal marine sediments accords with the findings of several other studies (e.g. Griffin and Goldberg, 1963; Parham, 1966; Keller, 1970).

The bulk of the sediments in the study area accumulated in an open marine shelf environment. The Taumarunui Formation flysch however was probably deposited in a deeper, localised, tectonic trough. In general, these marine sediments have non-diagnostic clay mineral assemblages. Within some formations however, there are lateral variations in clay mineral distribution that may relate to environmental conditions or to relative distance from sediment source areas.

Lateral variations in clay mineral species in the Taumarunui flysch sediments may be in part due to preferential flocculation and size settling of clay minerals upon entering the marine environment (cf. pp.116-119). This kind of data may in turn indicate directions of sediment transport. The relative abundance of kaolinite closest to source and of illite and chlorite further away parallels similar trends described by other workers (e.g. Powers, 1957; Parham, 1966; Gibbs, 1977). In the Taumatamaire siltstones kaolinite is only common in sediments adjacent to the basement Herangi High. The kaolinite was probably reworked from soil material on the High or from Waikato Coal Measures elevated by contemporaneous activity of the High (cf. p.116) and thus it reflects proximity of source. Sampling density within this formation was generally too sparse to permit any more authorative statement concerning the causes of lateral variation in clay mineral abundances.

With increasing distance from inferred source regions, the Mangarara and Purupuru sediments contain decreasing quantities of volcanic material and montmorillonite, and increasing abundance of illite and chlorite. Similar lateral variations in clay mineral abundance have been attributed to the marine alteration of montmorillonite to chlorite or illite (Grim and Johns, 1954; Johns and Grim, 1958; Milne and Early, 1958). In this case, however, the montmorillonite is considered to have been largely neoformed by the <u>in situ</u> marine alteration of volcanogenic materials and perhaps therefore, the higher montmorillonite concentrations closer to the source reflect an originally larger proportion of coarse-grained volcanogenic detritus in this area. Further from source the clay fraction may contain larger quantities of montmorillonite detrital from both volcanic and non-volcanic sources.

Certain clay mineralogic changes that begin in the depositional environment continue through to, and may be intensified in, the zone of diagenesis. For continuity of presentation examples of such changes are described in the next section.

3.5.5 DIAGENESIS

Diagenesis can significantly alter sediment porosity and permeability, through lithification and changes in mineral composition, and particle shape, size, surface texture and orientation (Krumbein, 1942). An appreciation of the changes produced during sediment diagenesis is important as these changes may significantly alter sediment clay mineralogy making it difficult to determine the clay minerals existing at the time of deposition, thereby limiting their use in paleoenvironmental interpretation. The following discussion, reviewing clay mineralogic changes that occur during sediment diagenesis, has been abstracted largely from Muller (1967) and Dunoyer de Segonzac (1970).

There is no universally accepted definition of diagenesis (e.g. Fairbridge, 1967; Muller, 1967; Dunoyer de Segonzac, 1970). In this thesis diagenesis is taken to include all physicochemical, biochemical and physical processes modifying sediments between their deposition and subsequent metamorphism. During diagenesis the clay mineralogic alterations begun prior to sediment deposition may continue during burial, and even be intensified. Diagenesis can be divided into several continuous stages, each with poorly defined limits, where the clay minerals may undergo continuous evolution (Dunoyer de Segonzac, 1970, p.333).

Early diagenesis occurs largely in the zone of Recent sedimentation. The upper limit is the sediment-water interface and the lower limit is where the sediment has lost about 50% of its connate waters and where lithification normally commences. Clay mineral assemblages may be complex in this environment as all species can be inherited. During early diagenesis clay minerals are essentially immobilised but circulating interstitial solutions carry dissolved substances which may cause clay mineral transformations and neoformations. Initially, trapped water within the sediment has much the same composition as the original fluid medium, but differences begin to appear as soon as the sediment is sealed from free circulation by deposition of later material. Changes in pH and Eh are accompanied by an increase in concentration of selective ions in solution. Such conditions promote clay mineral aggradations rather than degradations. Aggradational transformation commonly involves absorption of Mg^{2+} to form chlorite and/or chlorite mixed-layer clays, or K^+ to form illite and/or illite mixed-layer clays. Neoformations may be considerable, especially of montmorillonite in confined lacustrine and marine environments and by the marine alteration of volcanogenic materials.

Middle diagenesis is the zone where the sediments, although compacted, are still porous and permeable which allows circulation of waters of various origins (i.e. connate, meteoric, juvenile) to occur. The composition of these waters determines the kind and extent of dissolution, transformation and neoformation reactions that take place. As water is still present there is little dehydration and kaolinite and montmorillonite are stable. In more acid "siliceous" environments kaolinite is commonly formed. In more alkaline environments, rich in K^+ and Mg^{2+}, illite and chlorite are typically formed via mixed-layer clay mineral species.

Deep burial or late diagenesis is a more confined environment characterised by increasing pressure and temperature, more intense dissolution, and the compaction and expulsion of water. While compaction tends to expel part of the interstitial solutions, the clays, acting as osmotic filters, tend to concentrate ions. This contrasts with middle diagenesis where ions may be both leached and concentrated. During late diagenesis kaolinite and montmorillonite are unstable and clay mineral transformations always lead to crystalline illite and chlorite. The expulsion of water from the interlayers of clays under increasing pressures is an important process in the transformation of montmorillonite.

At several levels in the Cenozoic stratigraphic column beds of volcanic origin have been recognised that contain neoformed montmorillonite, and either α-cristobalite and/or analcite. It is worthwhile considering these mineral relationships in more detail in order to determine the time of formation and origin of these minerals. Numerous studies have shown that the neoformation of montmorillonite from volcanic materials can occur very rapidly (Norin, 1953; Morgenstein, 1967; Moberley _et al_., 1968) and perhaps even during sediment transport (Dunoyer de Segonzac, 1970, p.293). In the latter instance however the products are not post-burial diagenetic but rather halmyrolic (Muller, 1967). Experimental studies on glasses show that when immersed in water metallic cations are released from the glass surface in exchange for hydrogen ions in the water, making the water alkaline and increasing its capacity for dissolving silica. The rates of solution of commercial glass range from 1-100 mgm/100 cm^2 per day depending on water temperature, glass composition and water circulation (Hovestadt, 1902). Deffeyes (1959) estimated that natural glass shards of "the size range encountered in most sediments" would be expected to dissolve in 30-3000 years. If this is the case in seawater then montmorillonite formation from glassy material is most probably early diagenetic.

In the Purupuru Volcanic Sandstone montmorillonite formed prior to α-cristobalite cement, suggesting an early diagenetic origin for the montmorillonite (cf. pp.126-127). The reactions that take place to form montmorillonite are dependent on the chemical composition of both the parent material and the environment. In the marine environment however, seawater of 'normal' chemical composition has sufficient Mg^{2+} to support the formation of montmorillonite (cf. Harder, 1972). Where cristobalite is found coexisting with montmorillonite, it is considered that silica developed during glass dissolution and montmorillonite formation became excessive, assuming the constraints of the system were such that dissolved silica could not be flushed out. Periodically the silica would be forced to rapidly precipitate out of its highly saturated solution, perhaps as α-cristobalite. Hence α-cristobalite is considered a secondary product in the crystallisation of montmorillonite. More chemical work needs to be done to verify this suggested mineral formation sequence. Although only scarce analcite occurs in the Purupuru Volcanic Sandstones it is largely assumed, from mineral associations, to be authigenic.

Authigenic analcite coexists with neoformed montmorillonite in the Urenui tuffaceous sandstone. Analcite is commonly associated with sedimentary rocks (e.g. Keller, 1952; Norin, 1953; Deffeyes, 1959; Iijima and Utada, 1966) where it has been shown to form through the action of alkaline (Na-rich) pore waters on volcanic ash. The volcanogenic nature of the Urenui sandstone strongly suggests that this is a plausible origin. The mode of occurrence of the zeolites indicates they are authigenic and that they formed subsequent to montmorillonite (cf. p.133). Without further study it is difficult to say whether the alkaline environment necessary for analcite development is primarily a function of parent material composition or the result of the introduction of Na^+ from outside the system through circulating formational fluids.

It was earlier suggested that dissolved products from Oligocene soils combined in Oligocene sedimentary basins to neoform montmorillonite (cf. pp.105-111). The stage of diagenesis at which these reactions occurred cannot be determined exactly. However, the generally low rates of sedimentation in the Oligocene (Nelson, 1973), the fact that diagenetic intersititial solutions are usually more concentrated than seawater and the likely increases in pH and Mg^{2+} close below the sea floor (cf. pp.108-109), are all compatible with neoformation of montmorillonite in confined alkaline environments during early diagenesis.

Comparison of the clay mineralogy of early diagenetic concretions with those of their host mudstones indicates that the mudstone clay mineralogy approximates that existing at the time of, or shortly after, sediment deposition (Fig. 3.9). Similar comparisons could prove a useful indicator of clay mineral diagenetic changes in other sediments, particularly if the time of concretion formation is determined (cf. Curtis et al., 1972).

Comparison of Cenozoic clay mineral assemblages in sandstones with those in interbedded mudstones (cf. Fig. 3.8) strongly suggests that the clays in the sandstones have undergone degradational transformations, particularly as they are significantly less crystalline than clays in associated mudstones. Similarly the association of massive sandstones (e.g. Mt. Messenger and Omoao sandstones) with increased abundances of mixed-layer clay minerals, illite and degraded chlorite, together with a concomitant decrease in the abundance and crystallinity of montmorillonite suggests that alteration of clay minerals has been facilitated by the more permeable nature of these lithologies. The nature of the transformations is, however, difficult to determine as all recognised species of mixed-layer clays can occur in significant quantities in both mudstones and concretions (cf. Appendix Table IV.1), suggesting that a proportion of the mixed-layer clays in the sandstones may also be detrital.

The Purupuru and Urenui volcanogenic sandstones contain abundant montmorillonite in the clay fraction while the interbedded mudstones are characterised by illite, chlorite, montmorillonite and mixed-layer clays. The mineralogy contrasts markedly with the general clay mineralogic relationships determined for sandstones versus mudstones in this study (see previous discussion) and indeed, in other works (e.g. Weaver, 1958a; Keller, 1970). Grain fabric analysis shows that the volcanogenic sandstones are cemented by α-cristobalite and analcite that formed subsequent to the montmorillonite (Pls. 3.19 and 3.27). It is most probable that the early diagenetic montmorillonite was protected from diagenetic alteration by these cements which would have greatly reduced the permeability of the sandstones.

The association of abundant kaolinite and carbonaceous material in sandstones and mudstones of the Waikato Coal Measures and Maryville Coal Measures suggests the kaolinite may owe its origin, at least in part, to the degradational transformation of existing clays as a result of the circulation of dilute organic acids generated by the passage of pore waters through the carbonaceous lithologies (cf. Glass et al., 1955; Smoot, 1960; Keller, 1968b). The presence of highly degraded illite, chlorite, montmorillonite and mixed-layer clays in the sediments suggests the transformations affected mainly 2:1 clays that were detrital from adjacent sediments and soils (cf. pp.97-100 and pp.122-123), but part of the Waikato Coal Measures kaolinite possibly represents reconstitution of disordered kaolinite, detrital from soils developed on the adjacent Mesozoic basement rock landmass (cf. p. 100). The abundance of kaolinite in both mudstones and sandstones indicates that a sandstone environment is not a prerequisite for the diagenetic transformation of kaolinite. It may occur as long as the circulation of dilute acid solutions takes place. Rare neoformed kaolinite in Mangakotuku sediments (cf. Pl. 3.7) is compatible with localised acid leaching environments in the sediment (cf. Keller, 1968b), perhaps forming during the leaching of the scattered carbonate shells from the sediments during and/or soon after deposition (cf. Kear and Schofield, 1958; Nelson, 1973). Similarly the association of abundant kaolinite with poorly crystalline illite and chlorite in the banded Aotea sandstone (cf. pp.111-112) is compatible with an acid leaching environment, although a diagenetic origin for the clays has not been proven. Because the stability of kaolinite is very dependent on the chemistry of the environment, its distribution in the geologic column is not strictly related to temperature and pressure. For this reason kaolinite is a poor burial depth indicator (Dunoyer de Segonzac, 1970, pp.285-286).

The absence of direct mineralogic evidence as to the nature of the clay mineralogy of source materials for Mesozoic sediments allows only an estimate to be made of their existing content of diagenetic clays. Triassic sediments are characterised by largely illite, chlorite and mixed-layer clays (C-V) in the clay fraction. While their mineralogy could be detrital (cf. pp. 96-97) the absence of montmorillonite may be anticipated as estimates of depths of burial for the sediments of greater than 4 km (Kear 1960;

sandstones may differ because: (1) leaching by percolating ground waters in the more porous sandstone lithologies increases the kaolinite, illite and mixed-layer clay mineral abundance relative to other minerals, and (2) tuffaceous sandstones (e.g. Mangarara, Purupuru and Urenui sandstones) contain abundant volcanogenic montmorillonite which may be protected from diagenetic changes by α-cristobalite and analcite cement. Thus mudstone mineralogy tends to be indicative of provenance and the sedimentary environment while sandstone clay mineralogy is more indicative of post-depositional diagenetic effects.

On the basis of field and laboratory work in this study, together with some published data, the characteristic clay minerals of central Western North Island mudstones, summarised in Figures 3.5 and 3.6, were formed as follows:

The clay mineral suite in Mesozoic sediments is consistent with derivation from a western igneous/metamorphic provenance. However it is probable that the mixed-layer clays, and perhaps also some of the chlorite, have also developed by transformation during burial diagenesis. The kaolinite-rich Waikato Coal Measures clays are considered to have developed in part by prolonged chemical weathering of Mesozoic sediments and in part by diagenetic transformation under acid leaching conditions associated with carbonaceous sediments. The abundant crystalline montmorillonite in Oligocene Te Kuiti sediments is considered to have been primarily neoformed in shallow marine sedimentary basins characterised by very slow deposition and dominantly carbonate sedimentation. Crystalline illite and chlorite that characterise most of the Mahoenui to Nukumaruan (Miocene-Lower Pleistocene) sediments were mainly detrital from Mesozoic sediments under conditions of increased topographic relief, and rapid physical erosion as a result of accelerating tectonic activity during this period. The montmorillonite in Mahoenui and Mokau sediments (Lower Miocene) was largely detrital and derived from the erosion of uplifted Te Kuiti sediments. Montmorillonite in the Mohakatino sediments (Middle Miocene) was derived primarily from the alteration of volcanic products while much of the montmorillonite in Mt. Messenger to Tangahoe sediments (Upper Miocene to Pliocene) was derived from erosion of Mohakatino lithologies. The Plio-Pleistocene increase in montmorillonite abundance probably resulted from alteration of increasing quantities of volcanogenic material in these sediments derived from contemporaneous volcanism in the Central Volcanic and western Taranaki regions of the North Island.

Analysis of central Western North Island sediments shows that almost all kinds of clay minerals occur in the major depositional environments. Thus the potential value of predicting the environment of deposition from clay minerals is very dependent on differentiating detrital, transformed and neoformed clay minerals. Criteria found useful in distinguishing these clay mineral groups are summarised in Table 3.8. It is stressed that no one criterion is a diagnostic indicator and also that not all of these criteria will be applicable to other sediments.

The clay mineralogy, and indeed the bulk mineralogy, of Mesozoic to Pleistocene sediments of the central Western

RECOGNITION CRITERIA	DETRITAL CLAYS	TRANSFORMED CLAYS	NEOFORMED CLAYS
Commonly occurring clays	All clays	Kaolinite, chlorite and mixed-layer clays	Montmorillonite, illite. Dominance of one clay mineral species.
Clay mineral crystallinity	Low - high	Low-degradational transformations High-aggradational transformations	High
Sediment fabric	Randomly orientated aggregates and packets of clays	-	Interlocking plates
Clay particle morphology	Anhedral - euhedral	Anhedral - euhedral	Euhedralandlath-shaped grains and delicate structures. Montmorillonite forms waxy plates; kaolinite forms stacked hexagonal plate
Associated minerals	-	-	α-cristobalite, analcite, glauconite.
Characteristic lithology	None	None, but more common in porous sandstones	None

Table 3.3. Petrologic criteria used in this study to help distinguish between detrital, transformed and neoformed clay minerals.

North Island bears the imprint of a complex set of factors, including the influence of provenance, climate, tectonism, the depositional environment and diagenesis. In this clay petrologic study, as in many other fields of interpretive geology, it was frequently not possible to obtain direct proofs or unique interpretations. In many instances, however, sediments show significant variations in their clay mineral composition which can be related to different types and intensities of environmental conditions in the source and depositional areas.

Provenance was an important factor in determining the detrital nature of Miocene to Pleistocene sediments. These clays were primarily inherited from Mesozoic sediments, reworked from previously deposited Tertiary sediments, and/or neoformed from volcanogenic materials. A terrigenous source for clays was fostered by the readily erodible nature of the dominantly argillaceous Tertiary sediments.

Climate is considered to have been an important factor in determining the abundant kaolinite in the Waikato Coal Measures. The kaolinite is in part detrital, originating through prolonged (90 m.y.) deep chemical weathering of Mesozoic sediments under a warm temperate to subtropical climate on a tectonically stable landmass. In Oligocene times chemical weathering of the land surface was effected during a period of regional stability and warm temperate climate. These conditions resulted in minimal terrigenous input and the transport of largely dissolved products to the marine depositional basins where montmorillonite was primarily neoformed under a predominantly carbonate sedimentary regime. These relationships suggest that clay minerals are more likely to record the effects of climate when tectonism is slight.

Active tectonism is probably the single most important factor controlling the clay mineral composition of Miocene to Pleistocene sediments. Intermittent differential elevation and gentle tilting of sub-basin basement blocks between the major meridional growth fractures and their splays resulted in local downwarping of areas of deposition and elevation of areas bordering the depositional basins, including areas of previously deposited Tertiary sediments. Increased slopes and accelerated runoff generally hindered soil development, and mechanical erosion and weathering were dominant and sedimentary products detrital. Increasing tectonism is reflected by increasing quantities of crystalline illite and chlorite detrital from uplifted Mesozoic sediments. The character of the clay mineral assemblages released from Mesozoic sediment provenances and being deposited in the sedimentary basins is determined largely by the relative intensities of climate and tectonism in the source area (Fig. 3.20). Under active tectonics and cold or arid climates the clay minerals arriving in the depositional basin are overwhelmingly detrital and mirror source rock clay mineralogy. Under active tectonics and a warm humid climate there is extensive hydrolysis, moderate soil development and because of the active tectonic regime both the soils and underlying porous rocks are eroded and transported to the depositional basins. A warm humid climate and tectonic quiescence allows a thick soil profile to

<table>
<tr><td rowspan="2">TECTONISM</td><td>Tectonically quiescent
Low relief
Long time</td><td>Clays mainly detrital from parent rocks and soils; moderately crystalline illite and chlorite, and mixed-layer clays.</td><td>Clays detrital from soils; poorly crystalline illite, chlorite and mixed-layers. Clays neoformed in sediments and soils; crystalline montmorillonite.</td></tr>
<tr><td>Tectonically active
High relief
Short time</td><td>Clays solely detrital from parent rock; highly crystalline illite and chlorite.</td><td>Clays detrital from parent rock and soils; low to high crystallinity illite and chlorite, mixed-layer clays and kaolinite</td></tr>
<tr><td></td><td></td><td colspan="2">Temperate</td></tr>
<tr><td></td><td></td><td>Arid
and/or
Cold</td><td>Humid
and
Warm</td></tr>
<tr><td></td><td></td><td colspan="2">CLIMATE</td></tr>
</table>

Fig. 3.20. Diagram illustrating clay mineralogic end-members that are likely to be found in sediments when Mesozoic sandstone and mudstone source materials are subjected to varying climatic and tectonic intensities. Note that diagenesis may modify any of these clay mineral suites.

develop, the climate is conducive to growth of vegetation which further protects the thick weathered profile, dissolutions are considerable and highly degraded clays and dissolved products are transported to the sedimentary basins where montmorillonite may be neoformed. An arid or cold climate is generally unfavourable for soil development but tectonic quiescence allows more time for pedogenesis to take place.

No fundamental relationship exists between clay mineralogy and depositional environment in the study area. Any of the major species of clay mineral can occur in sediments formed in continental, marginal marine and shallow marine environments. Kaolinite is, however, most commonly associated with freshwater and marginal marine environments. Certain formations exhibit lateral changes in clay mineral abundance that are a function of the complex interaction of clay mineral flocculation and differential sedimentation and/or the differential sedimentation of non-clay volcanogenic material that is later altered to clay minerals.

Diagenesis played an important role in the formation of certain clay mineral assemblages. Montmorillonite is the early diagenetic product of the alteration of glassy pyroclastic material in several formations. The early diagenetic formation of montmorillonite in Oligocene sedimentary basins is an example of neoformation of clay minerals under extremely low rates of sedimentation where sufficient time prevailed for chemical equilibrium to be achieved between incoming dissolved products and the environment. During shallow burial diagenesis kaolinite formed, or was reconstituted, under the influence of circulating dilute acid solutions associated with carbonaceous material in Waikato Coal Measures and Maryville Coal Measures sediments. A number of the more porous Tertiary sandstones contain kaolinite and mixed-layer clays formed by the degradational alteration of pre-existing clays by formational fluids. The abundant chlorite and chlorite mixed-layer clays in Mesozoic and particularly Triassic sediments may in part reflect progressive transformation of montmorillonite during deep burial diagenesis.

In conclusion, this study emphasises that no definitive relationship exists between clay mineralogy and depositional environment. Moreover, there is no single criterion that can be used to designate a clay mineral as detrital, transformed or neoformed, which are necessary prerequisites before any environmental assessment can be attempted. The distinction may only be established using collectively the evidence of clay composition, morphology, structure and distribution, and by considering these factors in conjunction with the bulk petrologic properties of the sediment. Clay minerals are useful for provenance and paleoenvironmental interpretation only in favourable cases.

CHAPTER 4

CLAY PETROLOGY OF SURFICIAL WESTERN SHELF SEDIMENTS

4.1 INTRODUCTION

This chapter describes the mineralogy, in particular the clay mineralogy, of the surficial sediments on that part of the western continental shelf of the North Island between the Manukau Harbour in the north and the mouth of the Rangitikei River in the south, and seaward to about the 175 m isobath, a total area of some 45,000 km^2 (Figs. 1.1 and 4.1).

4.1.1 WESTERN SHELF MORPHOLOGY

For the purpose of this discussion the western shelf is subdivided into the Hamilton, North Taranaki, Central Taranaki and South Taranaki Shelves, and the North Cook Strait Basin (Fig. 4.1); the last named area refers to the northern part of the Cook Strait Basin defined by Lewis and Eade (1974).

Seaward of Manukau Harbour the outer edge of the continental shelf is sharply defined some 40 km offshore. Further to the south, between Manukau Harbour and Cape Egmont, the shelf widens to 95 km and the shelf-slope break is less definite. South of Cape Egmont the continental shelf continues to widen and links with that portion of the shelf running northeast from Cape Farewell (northwestern South Island). Morphologically, the continental shelf consists of two parts: (1) the Cook Strait Basin, and

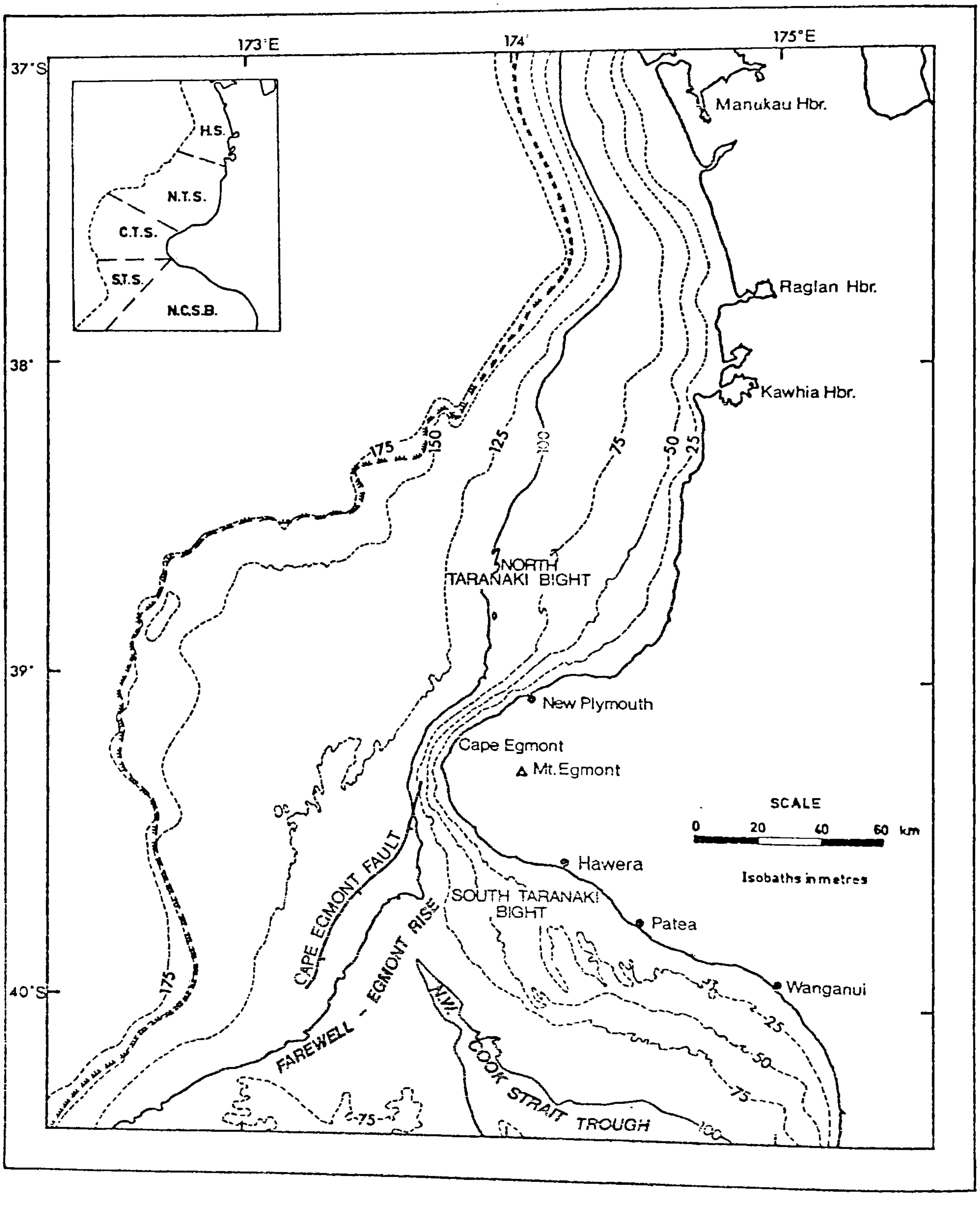


Fig. 4.1. Bathymetry and main physiographic features of the western continental shelf of the central Western North Island. Inset map shows the division of the western shelf into the Hamilton Shelf (H.S.), North Taranaki Shelf (N.T.S.), Central Taranaki Shelf (C.T.S.), South Taranaki Shelf (S.T.S.) and North Cook Strait Basin (N.C.S.B.). Edge of continental shelf marked by ⊥⊥⊥⊥⊥. (After McDougall and Brodie, 1967; Lewis and Eade, 1974). Additional "lift-out" copy available in back pocket.

(2) the western shelf proper; these are separated by the Farewell-Egmont Rise (Lewis and Eade, 1974, p.1).

The western shelf is a smooth submarine plain sloping seawards about $0^{o}17'$ between Manukau Harbour and Raglan Harbour and decreasing to $0^{o}4'$ between Raglan Harbour and Cape Egmont. Local topographic irregularities occur off Cape Egmont where volcanics occur on the shelf, in the vicinity of several rocky outcrops off Patea, and adjacent to the Farewell-Egmont Rise (where the Cape Egmont Fault forms a low submarine ridge about 3 m high). West of Cape Egmont, the presence of marked irregularities in the 125 m isobath (Figs. 4.1) and strong subsurface reflectors indicate the presence of a submerged and largely buried shoreline named the Egmont Gulf (McDougall and Brodie, 1967, p.42; cf. Fig. 4.1).

The coastline morphology varies considerably throughout the study area. From the Manukau Harbour to Raglan Harbour narrow sandy beaches give way to subdued cliffs of Tertiary and Quaternary strata. The coastline south from Raglan Harbour is characterised by the presence of three large shallow harbours (Raglan, Aotea and Kawhia Harbours) and narrow, sandy beaches backed by large dunes. Extending south from Kawhia Harbour to just north of New Plymouth are very narrow sandy beaches with cliffs up to 240 m high cut in Mesozoic and Upper Miocene sediments, or raised coastal terraces with fossil dunes. From the Mimi River south to Hawera gravelly beaches and sandy inlets are backed by cliffs averaging 30 m high in fossil dunes and andesitic lahar flows. From this point south to the Rangitikei River very narrow sandy beaches give way to cliffs up to 45 m high cut in coastal terraces with fossil and active dunes.

Harbours, rivers and streams entering the sea drain a wide variety of lithologies (Fig. 1.3).

4.1.2 WESTERN SHELF SEDIMENTOLOGY

In the southern part of the study area, oceanic circulation is governed by the northeasterly flowing Westland Current, a branch of which enters Cook Strait as the D'Urville Current (Fig. 4.2). In the north, the southerly flowing West Auckland Current probably results in a zone of mixing extending as far south as Raglan Harbour (Brodie, 1960).

Sediments of the western shelf are mainly muddy fine sands, muddy very fine sands, and sandy muds (Fig. 4.3). North of Cape Egmont, the fine sand zone in the middle shelf, unlike the outer and inner very fine sand zones, contains material up to very coarse sand and granule size. Sediments of the Central Mud Belt contain less than 10% sand. In general, much of the coarser material consists of calcareous biogenic fragments. The coarser sediment zones adjacent to Cape Egmont and in the northern North Cook Strait Basin are mainly slightly gravelly and gravelly medium to very coarse sands, the former area containing numerous andesite cobbles and boulders up to 10 km from shore. At the shelf edge and on the upper slope, sediments consist mainly of molluscan, bryozoan and benthic foraminiferal debris. Further downslope, bioclasts become smaller and sediments are typically foraminiferal and coccolith oozes (McDougall, 1972; Carter 1975, p.230).

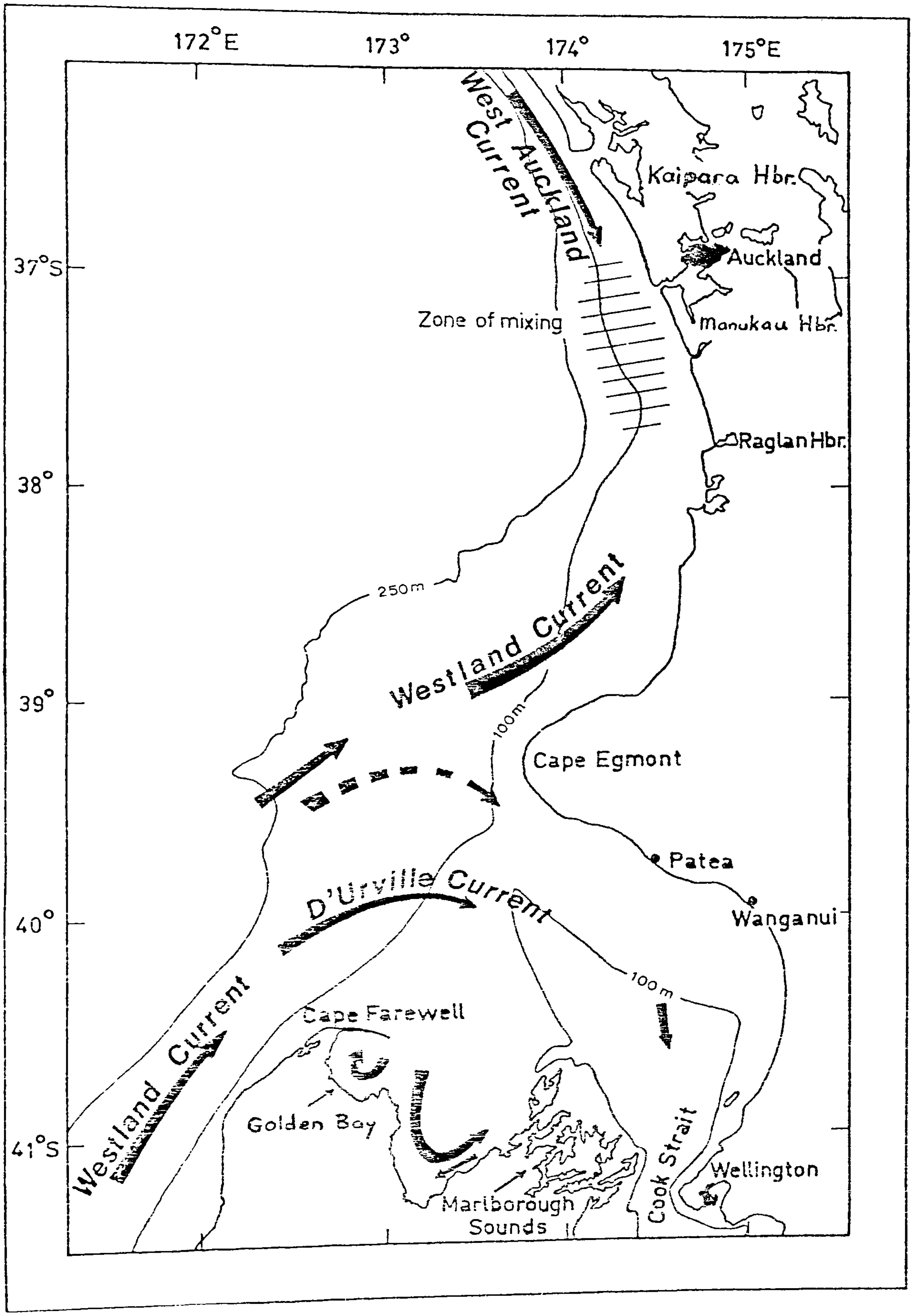


Fig. 4.2. Major currents and circulation patterns over the western continental shelf and adjacent shelf areas (after Brodie, 1960; Heath, 1969; Heath, 1974).

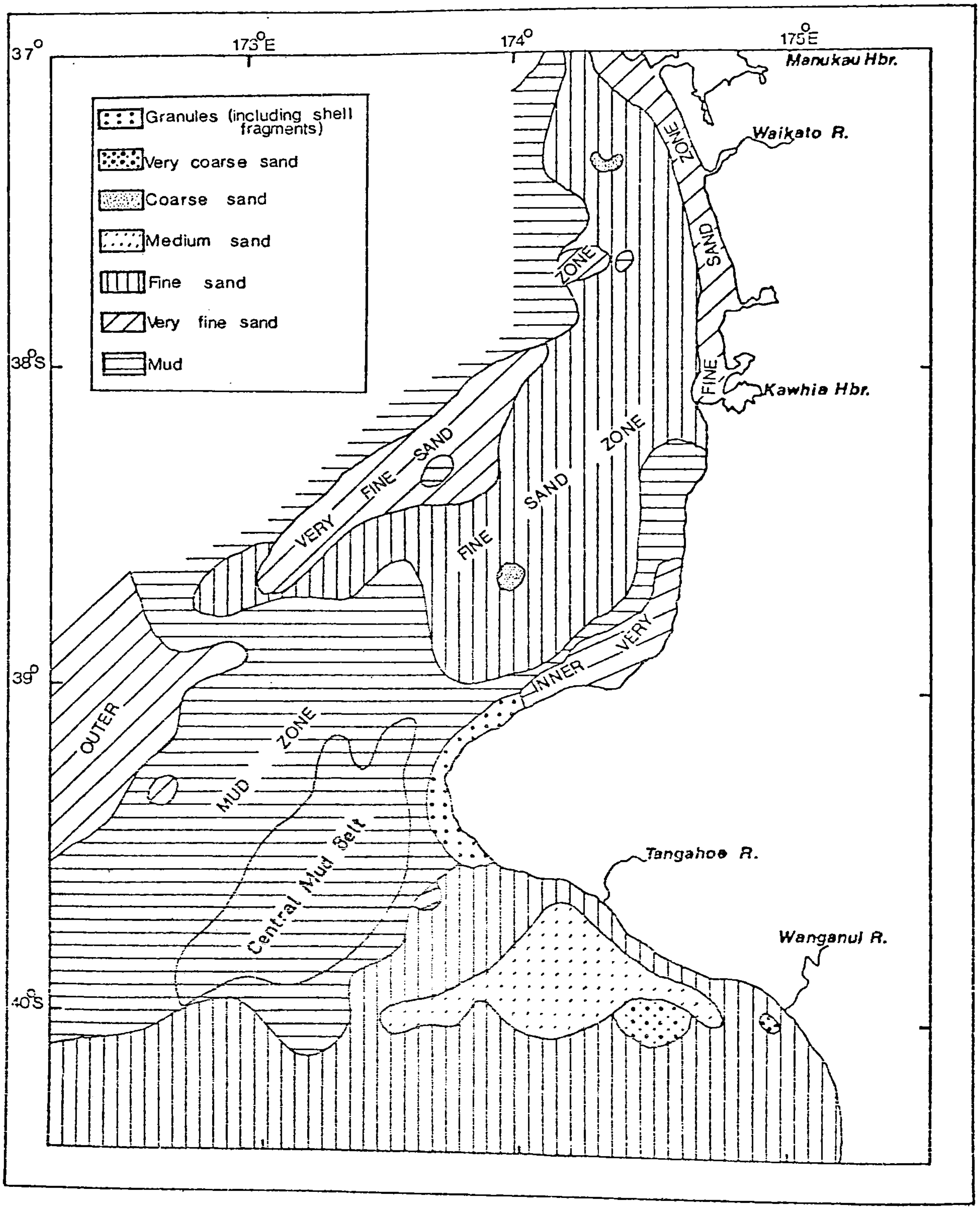


Fig. 4.3. Distribution of modal sizes in the sediments of the western continental shelf (after McDougall and Brodie, 1967; Lewis and Eade, 1974). Additional "lift-out" copy available in back pocket.

The present distribution pattern of sediments on the shelf does not fit the classical picture of progressive fining away from shore in response to the decreasing environmental energy of deepening waters. North of Cape Egmont, the sediments are characterised by a zone of fine sands flanked by very fine sands and muds. The central zone of coarser sands, and possibly the outer very fine sand zone, are presumed to represent relict sediments deposited during the Last Glacial maximum, about 15,000 - 20,000 yrs B.P., and/or during the subsequent Aranuian marine transgression (McDougall and Brodie, 1967).

The present distribution pattern of surface sediments on the western shelf was initiated during the last stadial, about 20,000 yrs ago, when sealevel was some 120 m below its present level (Lewis and Eade, 1974, p.5). At that time, the Farewell-Egmont Rise (Fig. 4.1) probably formed a continuous land bridge between the North and South Islands and shorelines lay between the present 100 m and 125 m isobaths. Sand and mud from the glacial rivers of Westland were moved north along this ancient coast by southwesterly swells and the Westland Current to the continental shelf and slope off Taranaki. The resulting deposits were mainly shelly sands, as the very shallow nature of the shelf (nowhere deeper than 55 m) enabled storm waves to keep mud in suspension until it moved beyond what is now the 150-175 m shelf break. At this stage, most of the Cook Strait Basin east of the Farewell-Egmont Rise was dry land, with sea restricted to a narrow trough extending from the present Cook Strait northwest towards the Rise. During the Last

Glacial maximum, the land bridge would have stood about 25 m above sealevel and was some 40 km wide at its lowest and narrowest place. A shallow depression, the Egmont Gulf, opened to the north from this saddle (Lewis and Eade, 1974).

During the Post-Glacial rise of sealevel about 15,000 yrs ago, the sea breached the Farewell-Egmont Rise at the saddle to link the Tasman Sea and the Pacific Ocean. From 15,000 - 10,000 yrs B.P. the sea covered most of the Egmont-Farewell Rise and sands, now represented by relict sediments north of Cape Egmont and in the North Cook Strait, were deposited on most of the present middle shelf. The present regime of deposition commenced about 6,000 yrs ago when the sea rose to within a few metres of its present level (Lewis and Eade, 1974).

Contemporary sediments are derived from river transported material, from cliff-line erosion, from biogenic material, and from relict sediments reworked by currents, wave action and burrowing organisms. Modern sediments are principally terrigenous muds and very fine sands that are being deposited in a coastal belt some 10-30 km wide, and in local areas free of strong currents (McDougall and Brodie, 1967). The interaction of the D'Urville Current (Fig. 4.2) with tidal and westerly flowing wind-generated currents leaves areas of comparatively quiet water in the region of the Central Mud Belt and in the centre of the Cook Strait Basin (Figs. 4.1 and 4.3). Mud-laden water from Western rivers appears to flow rapidly into Cook Strait. Some of the mud is carried directly through Cook Strait to the

Pacific Ocean, but most is transported by each of the two branch currents and deposited in the relatively quiet Central Mud Belt (Fig. 4.3) and the Mud Zone of the central Cook Strait Basin (Lewis and Eade, 1974). In the absence of strong currents, sediment resuspended by storm swells resettles in the same area. At other places on the middle shelf, mud deposition is inhibited by boundary layer turbulence associated with coarse-grained relict sediments. Any mud settling in these regions is probably removed by currents the next time it is thrown into suspension by bottom stirring storm swells. Although small amounts of mud may reach the outer-shelf, deposition is largely prevented by turbulence associated with the interaction of tides and ocean currents and the sediment is by-passed to the continental slope and beyond (Van der Linden, 1969, p.102).

Calcareous, and to a lesser extent siliceous, biogenic material forms a significant part of western shelf sediments in several areas. There is little detailed information available on the distribution of the major biogenic components in the sediments, which include bivalve and gastropod valves, bryozoan fragments, foraminiferal and coccolith tests, and diatom fragments. The distribution is complicated by the fact that in some areas biogenic deposits are a mixture of both relict and modern specimens. In general, an abundance of biogenic material implies a dearth of terrigenous sediment (Carter, 1975).

4.2 COMPOSITION OF SEDIMENTS

The composition of western continental shelf and adjacent river and harbour sediments is summarised on mineral concentration isoline diagrams (Figs. 4.4 - 4.15). The concentration of each major mineral species in river and harbour sediments is given in the oval adjacent to each river and harbour. Sampling sites (cf. Fig. 2.2) are shown by the large dots. The diagrams were constructed by drawing lines of equal mineral abundance about mineral concentration spot values. The reliability of the resulting mineral distribution patterns are dependent on the density of spot values. This, in turn, limits the amount of interpretation that can be placed on the data. For these reasons the isoline diagrams show only general trends in mineral abundances over the shelf. For example, Fig. 4.15 shows that montmorillonite is abundant in the clay fraction of Hamilton shelf sediments, common in Central Taranaki Shelf sediments and is relatively scarce in North Taranaki Shelf, South Taranaki Shelf and North Cook Strait Basin sediments.

2.2.1 BULK MINERALOGY

The calcium carbonate fraction of the western shelf sediments consists almost exclusively of biogenic material (Fig. 4.4). High concentrations of calcium carbonate west of the Mokau River mouth and south of Cape Egmont result from the widespread occurrence of skeletal material (dominantly molluscan and bryzoan fragments) in these areas (McDougall and Brodie, 1967). Outer shelf concentrations of carbonate material west of Raglan Harbour and west and southwest of Cape Egmont contain numerous foraminiferal and/or coccolith shells (Pl. 4.1), which attest to low rates of terrigenous sedimentation in these areas.

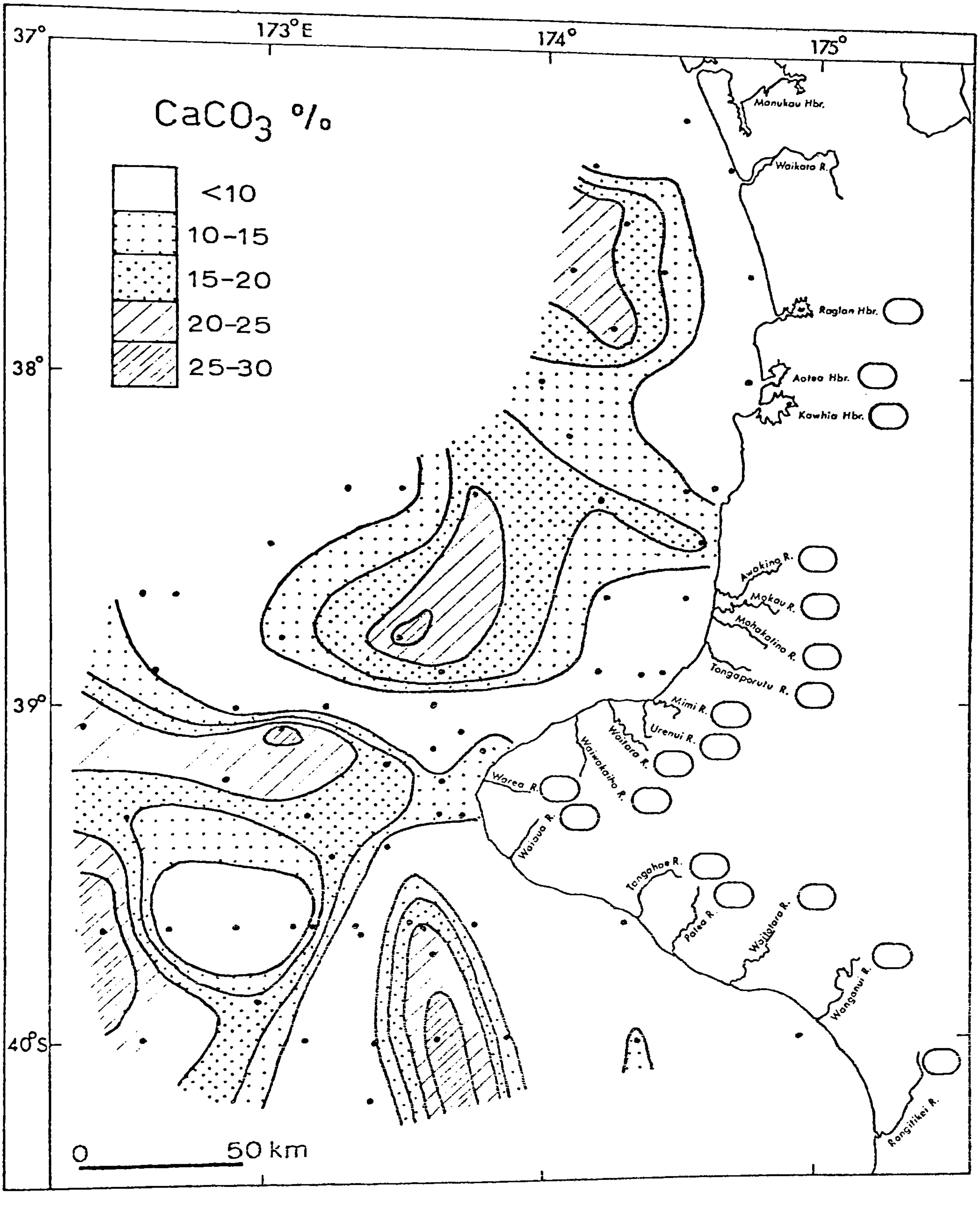


Fig. 4.4. Distribution of $CaCO_3$ in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

Siliceous biogenic material also occurs in shelf sediments, particularly as diatom tests (Pl. 4.1), but optical, TEM and SEM examination of samples indicate the total abundance of this material is low.

The abundance of components in the non-carbonate fraction was determined using XRD modal analysis and can be assessed from Fig. 4.4. Quartz, plagioclase feldspar, and crystalline clay minerals are the most abundant constituents, but "other" components, including amorphous material (mostly volcanic glass and allophanic clay), opaque and ferromagnesian minerals, potash feldspars, and siliceous organisms are locally common. The distribution patterns of the major non-carbonate components vary widely and in many instances mineral abundances offshore bear little resemblance to those being delivered from adjacent rivers and harbours. This undoubtedly reflects the fact that western shelf sediments are the product of a variety of processes including shoreline erosion, fluvial input, mineral authigenesis, and the effects of past and present hydrodynamic regimes.

To the north and west of Cape Egmont, quartz abundance (Fig. 4.5) tends to increase seawards, but south of Cape Egmont quartz is most abundant in a middle shelf position. Although plagioclase tends to be most abundant over the eastern portions of the shelf (Fig. 4.6) and to the extreme southwest, its distribution is less regular than that of quartz. The plagioclase feldspar component of most samples is characterised by the presence of both the A and B species (cf. pp.26-27). In general, quartz and feldspar abundances show little quantitative similarity to adjacent river and

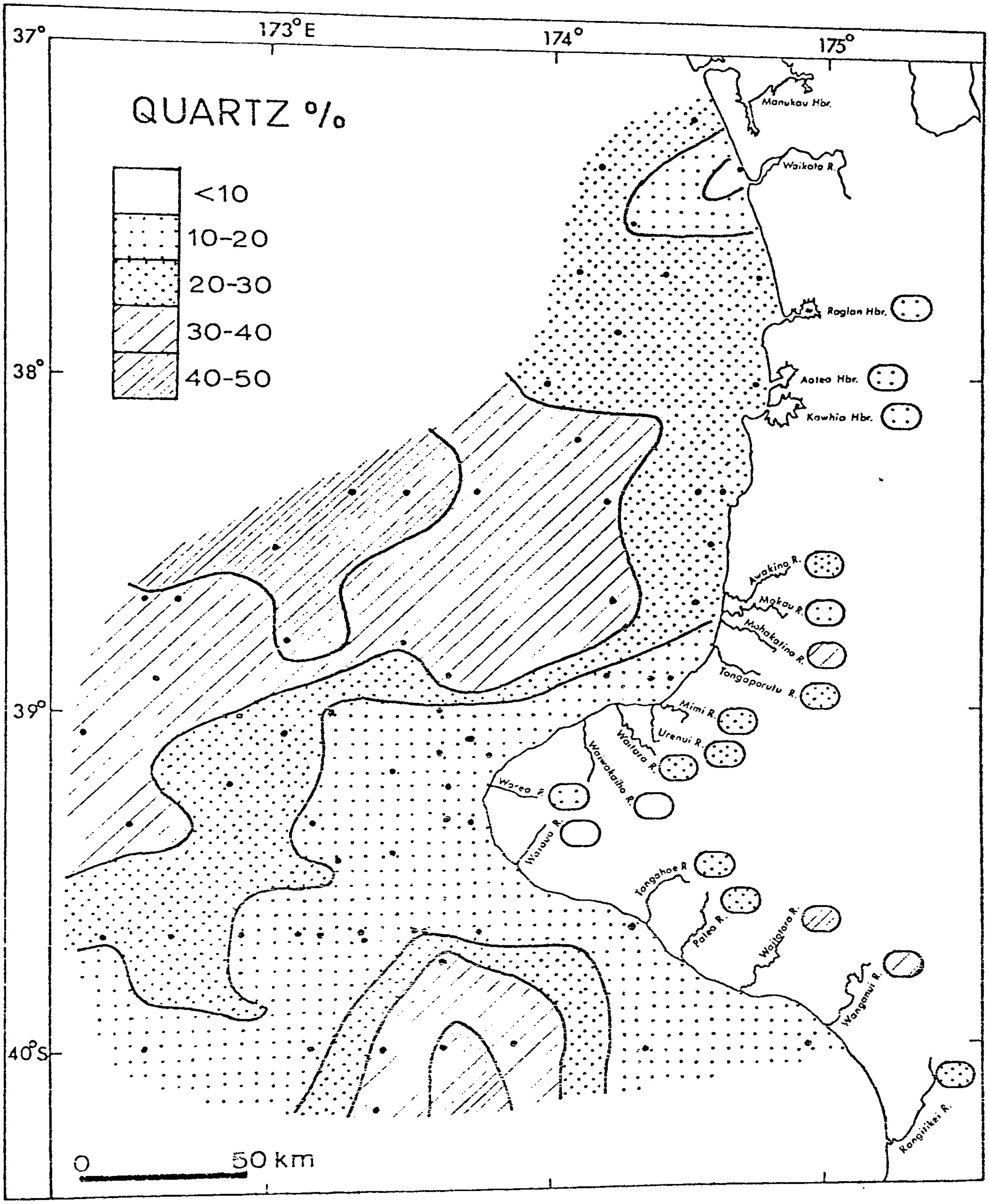


Fig. 4.5. Distribution of quartz (percent in acid insoluble residue) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

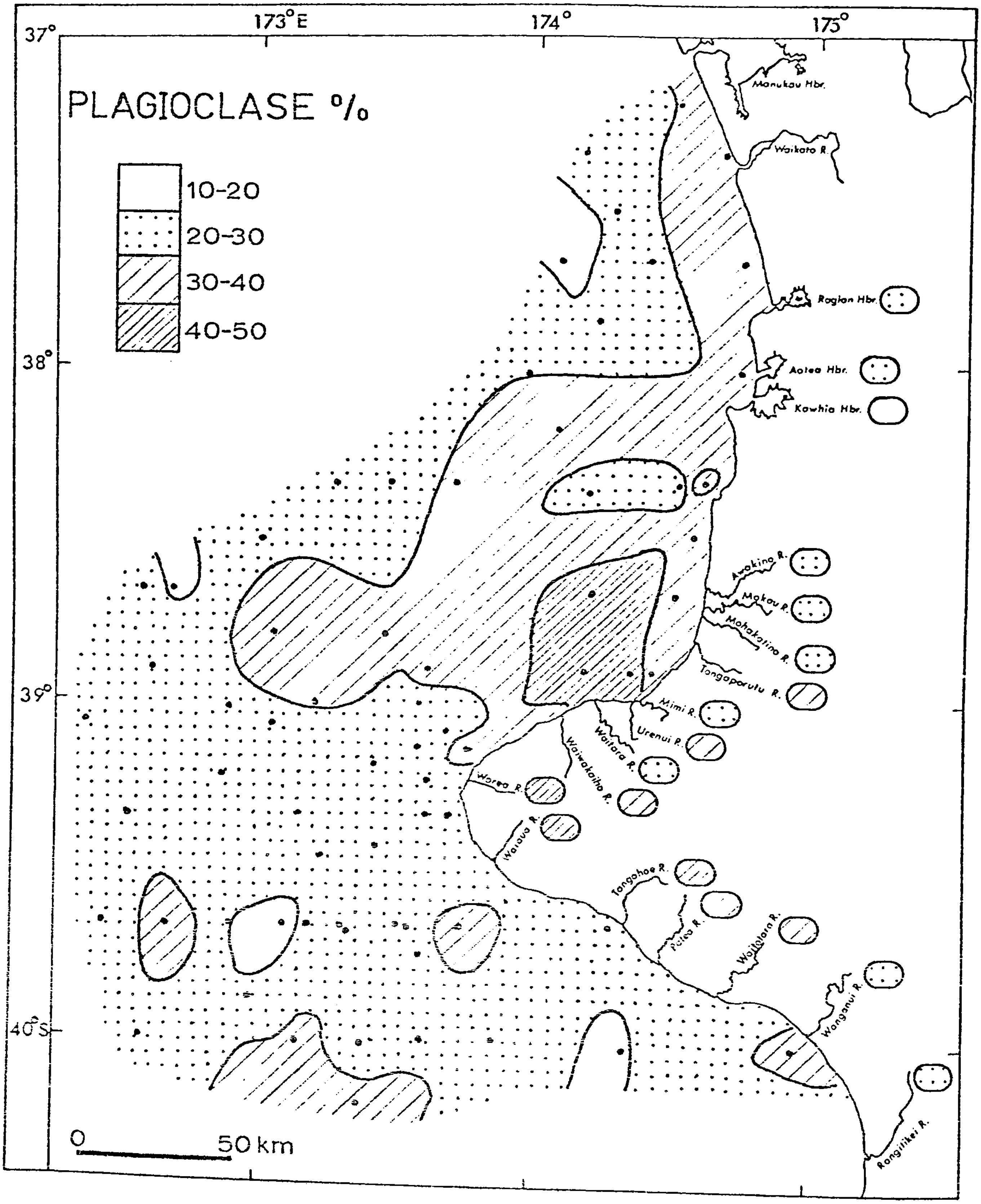


Fig. 4.6. Distribution of plagioclase feldspar (percent in acid insoluble residue) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

harbour sediments.

Potash feldspar constitutes less than about 10% of most samples and shows an irregular distribution pattern.

The total clay mineral distribution pattern (Fig. 4.7) follows only partly the distribution of mud size material (Fig. 4.3) on the shelf. Mud is concentrated in outer and inner zones north of Cape Egmont and in a broad southwesterly trending zone off Cape Egmont. Crystalline clay minerals are most abundant in sediments of the Taranaki Shelf and North Cook Strait Basin. Clay mineral concentrations are localised in two groups within the Mud Zone. One lies in a central shelf position west of Cape Egmont, the other constitutes the southwestern part of the Mud Zone and lies parallel to the Farewell-Egmont Rise with its north-eastern lobe extending through the saddle. In general the abundance of total crystalline clay minerals in river and harbour sediments bears little resemblance to that in adjacent western shelf sediments.

"Other" constituents are concentrated in nearshore areas, west of Raglan Harbour, and in a few isolated middle and outer shelf positions (Fig. 4.8). Opaque minerals (mostly titanomagnetite) and andesitic rock fragments constitute the bulk of the other material at two nearshore sites (Fig. 2.2 : B667 and B653). The distribution of titanomagnetite in western shelf sediments has been described by McDougall (1961). At the remaining locations (Fig. 4.8), "other material" consists largely of clay-sized, amorphous, siliceous material, much of which could be diatom fragments (c.f. Pl. 4.1), and occasional volcanic glass. Concentrations

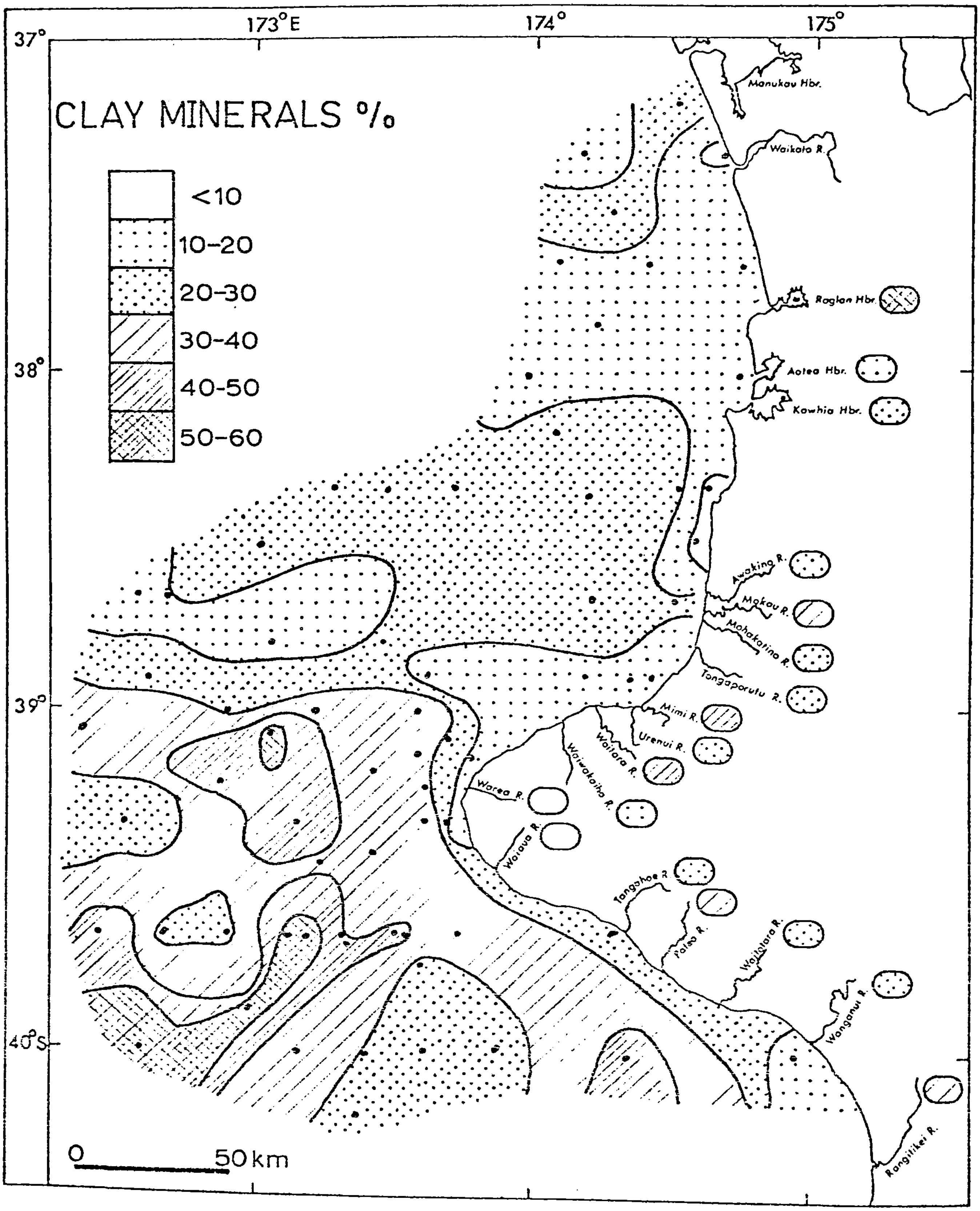


Fig. 4.7. Distribution of clay minerals (percent in acid insoluble residue) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

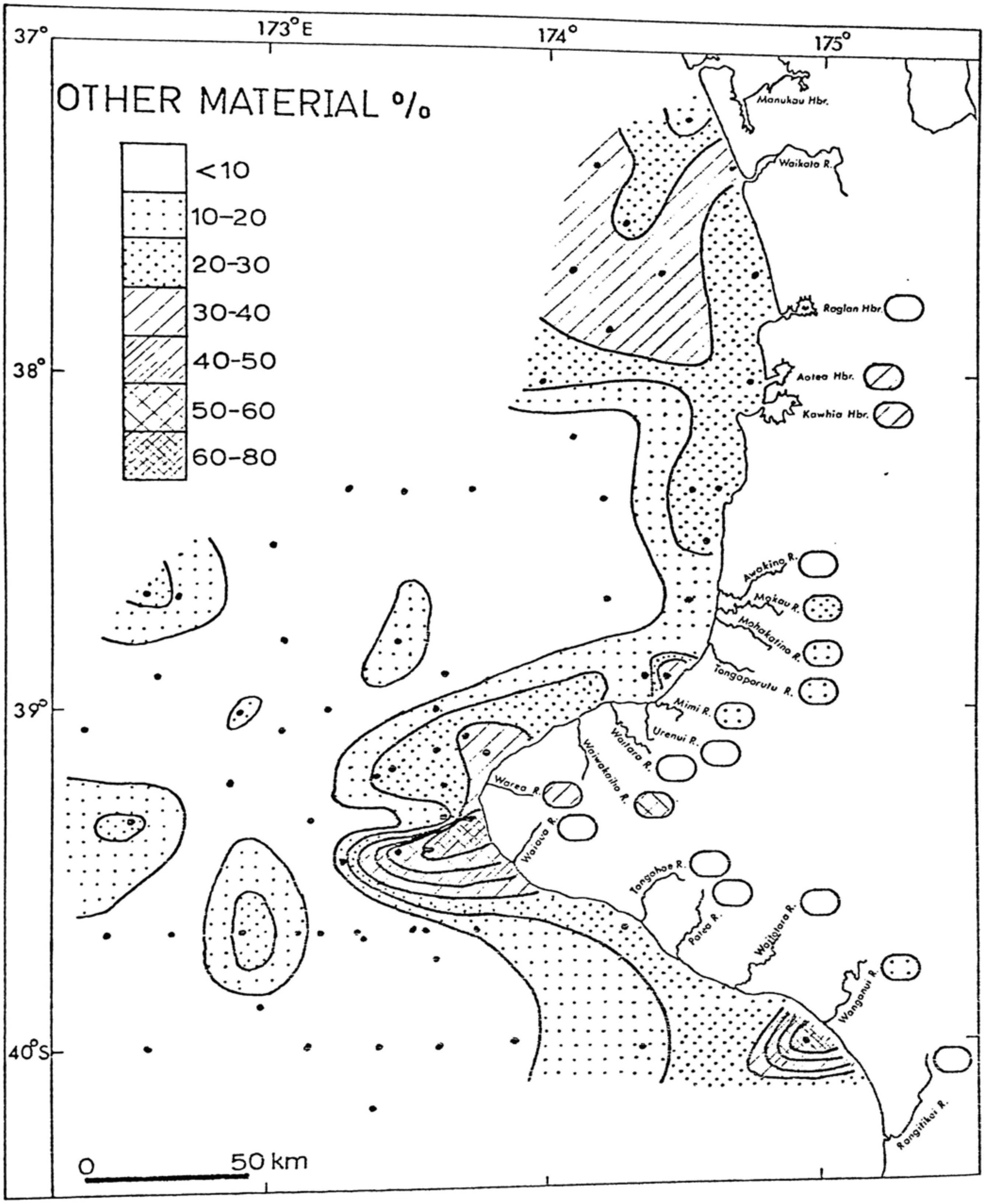


Fig. 4.8. Distribution of "other material" (percent in acid insoluble residue in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

of X-ray amorphous material west of Raglan Harbour and the Waikato River mouth lie in an area now occupied largely by relict sediments (McDougall and Brodie, 1967).

4.2.2 CLAY MINERALOGY

Western Shelf sediments contain mostly illite and montmorillonite, lesser amounts of chlorite and mixed-layer clays, and minor kaolinite. Trace amounts of halloysite are present in a few samples.

4.2.2A KAOLINITE

Kaolinite is rare or absent south of Cape Egmont but consistently forms some 5 to 10% of the clay mineral fraction in sediments to the north of the Cape (Fig. 4.9). The distribution of kaolinite in river and harbour sediments adjacent to the shelf shows a parallel trend. Kaolinite is most abundant in sediments from Raglan Harbour and the Mokau and Mohakatino Rivers.

4.2.2B ILLITE

Overall, illite is the dominant clay mineral in western shelf sediments. It is most abundant in sediments of the North Taranaki Shelf, South Taranaki Shelf and North Cook Strait Basin (Fig. 4.10). Crystallinity decreases towards the north (Fig. 4.11). The trends in both illite abundance and crystallinity are paralleled by generally similar trends in these properties for illite in the adjacent river and harbour sediments and in the onshore sediment formations.

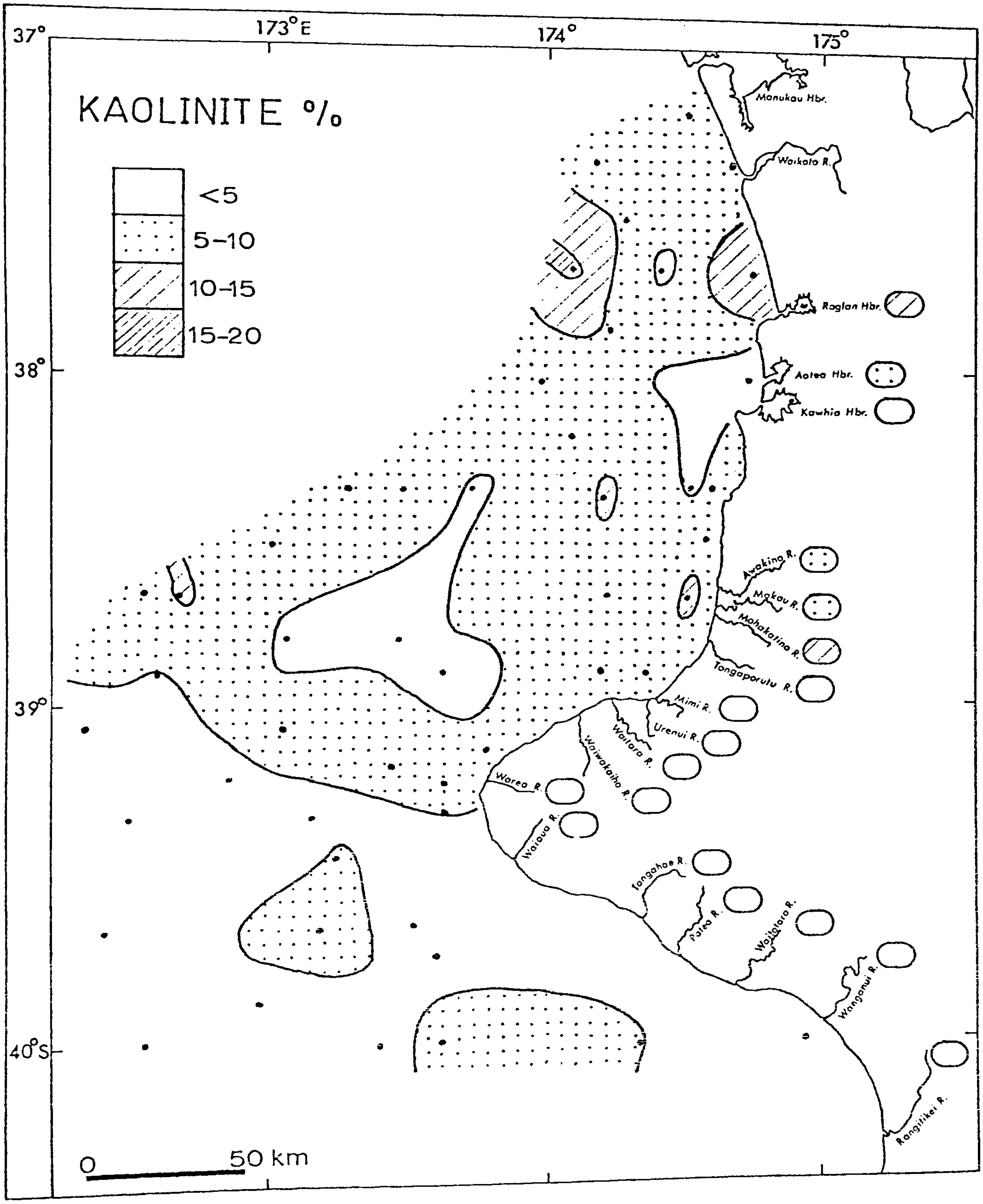


Fig. 4.9. Distribution of kaolinite (percentage of <2μ size fractions) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

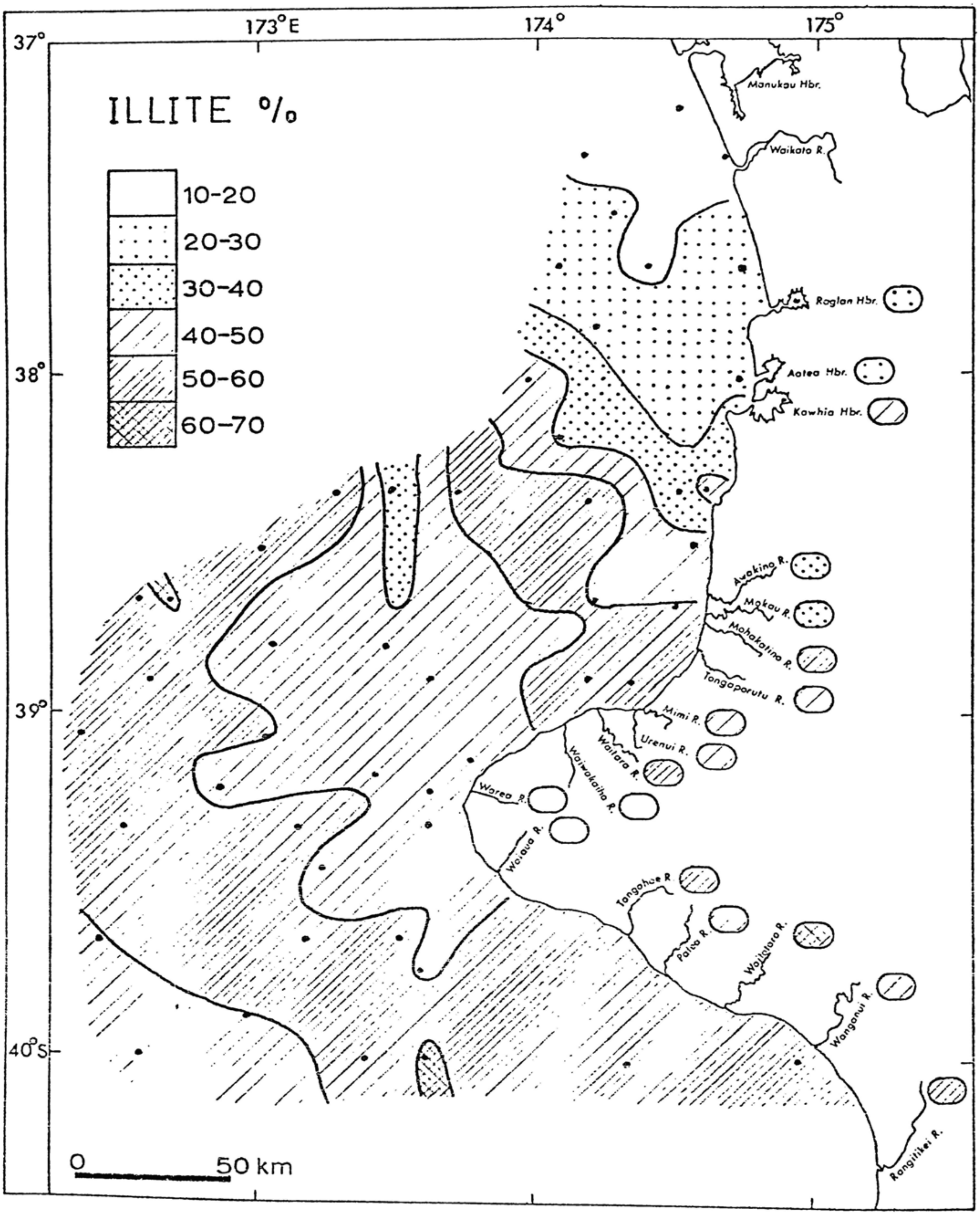


Fig. 4.10. Distribution of illite (percentage of $<2\mu$ size fractions) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

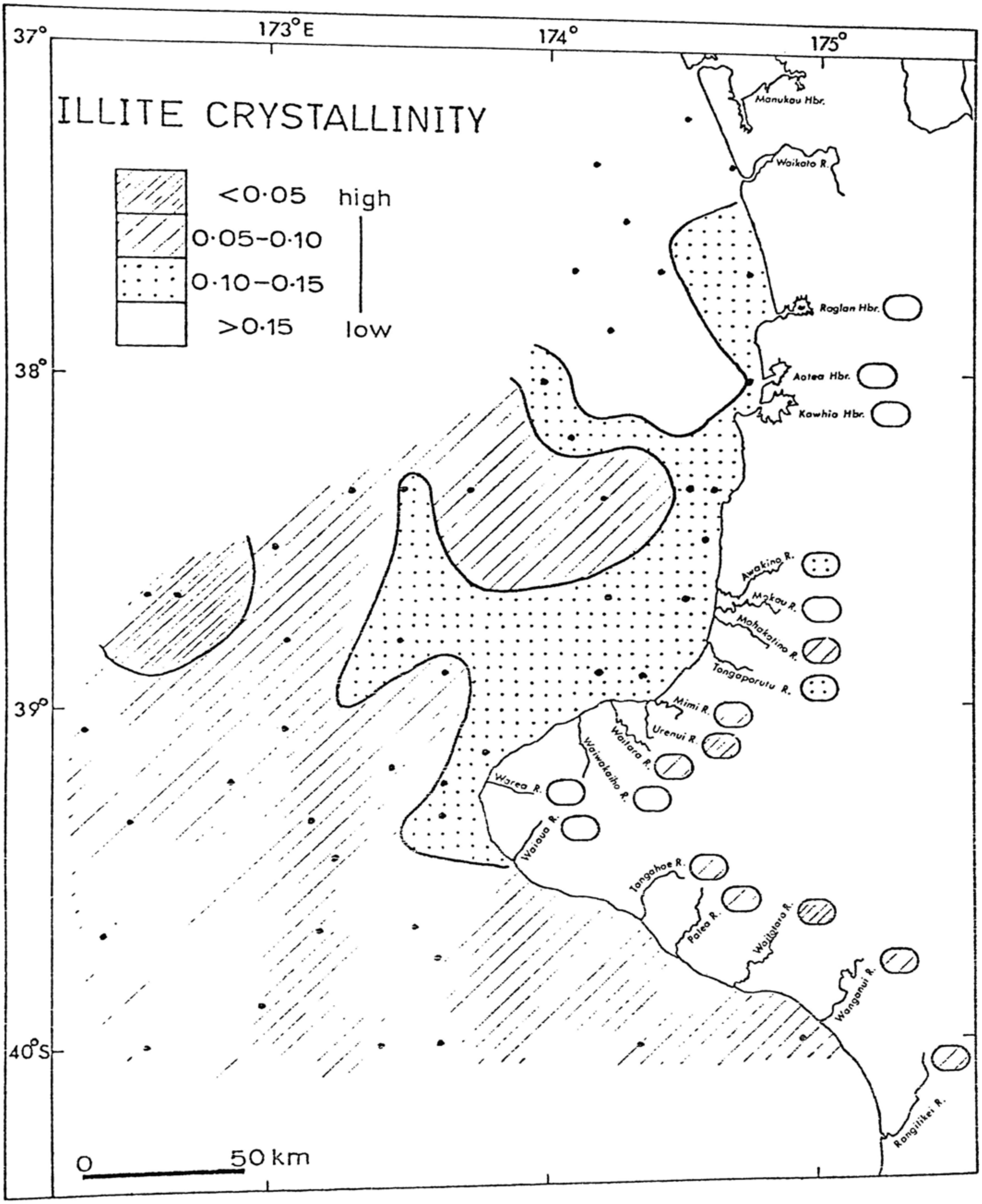


Fig. 4.11. Crystallinity of illite in $<2\mu$ size fraction of surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

4.2.2C CHLORITE

Although less regular, the distribution pattern for chlorite is similar to that for illite, with both chlorite abundance and crystallinity decreasing northwards (Figs. 4.12 and 4.13). Chlorite is most abundant and crystalline in sediments from the southwestern part of the shelf. These general trends are matched in part by trends in the abundance and degree of crystallinity of chlorite in the clay fraction of adjacent river and harbour sediments.

4.2.2D MIXED-LAYER CLAYS

Mixed-layer clays are most abundant south of Cape Egmont and in the North Taranaki Shelf sediments, but overall form on irregular distribution pattern (Fig. 4.14). Both the western shelf and the river and harbour sediments contain several different species of mixed-layer clay minerals (Appendix Table IV.1), the most common of which are C-V and C-M with lesser amounts of C-I. The frequency distribution of individual mixed-layer clay mineral species in both shelf and harbour sediments is random so that these minerals appear to be of little use as provenance indicators.

4.2.2E MONTMORILLONITE

Montmorillonite is abundant in the clay fraction of Hamilton Shelf sediments, common in Central Taranaki Shelf sediments and is relatively scarce in sediments of the North Taranaki Shelf, South Taranaki Shelf and North Cook Strait Basin (Fig. 4.15). The systematic increase in abundance of montmorillonite in sediments north of Cape Egmont is paralleled by a similar trend in the sediments of adjacent

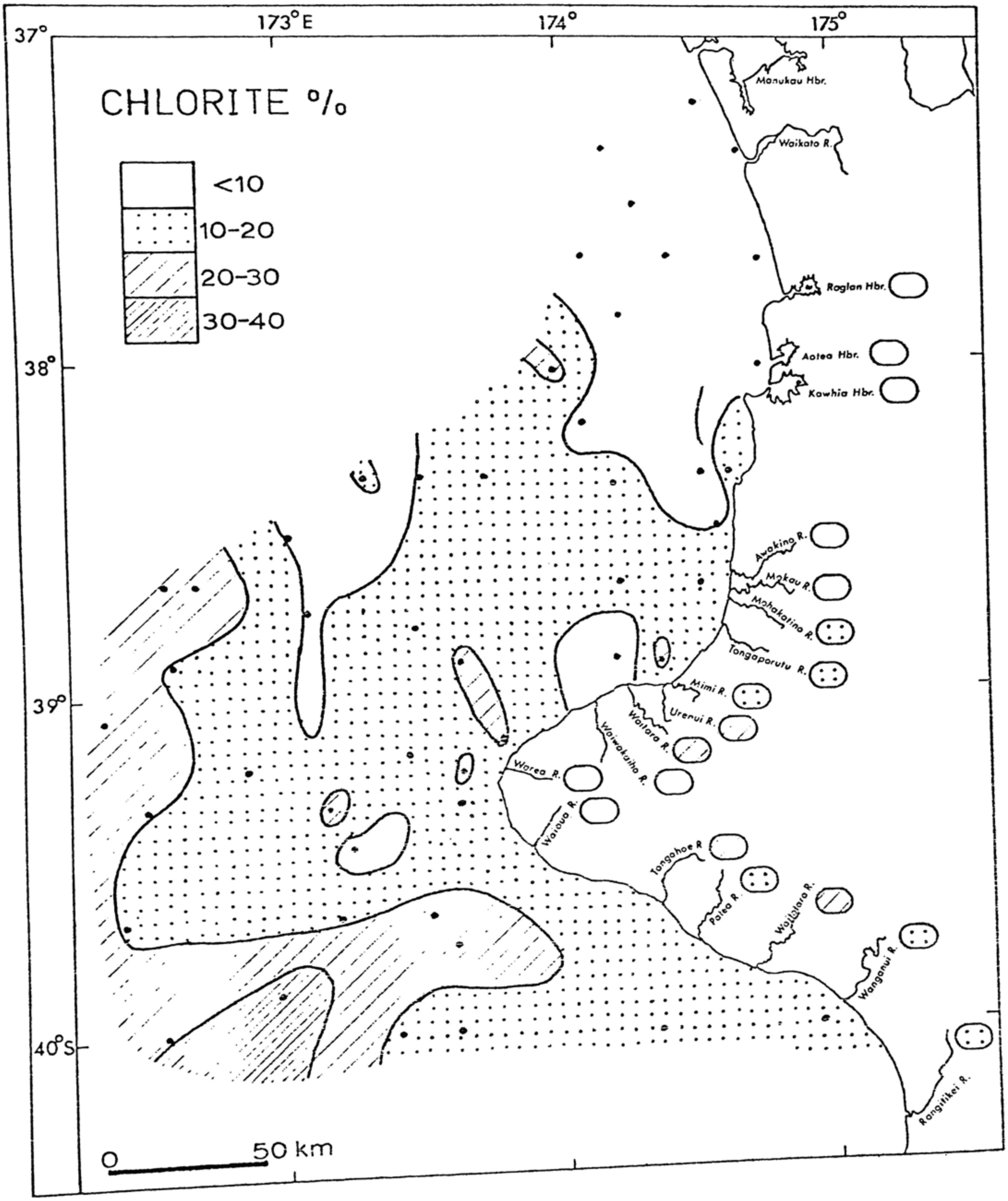


Fig. 4.12. Distribution of chlorite (percentage of <2μ size fractions) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

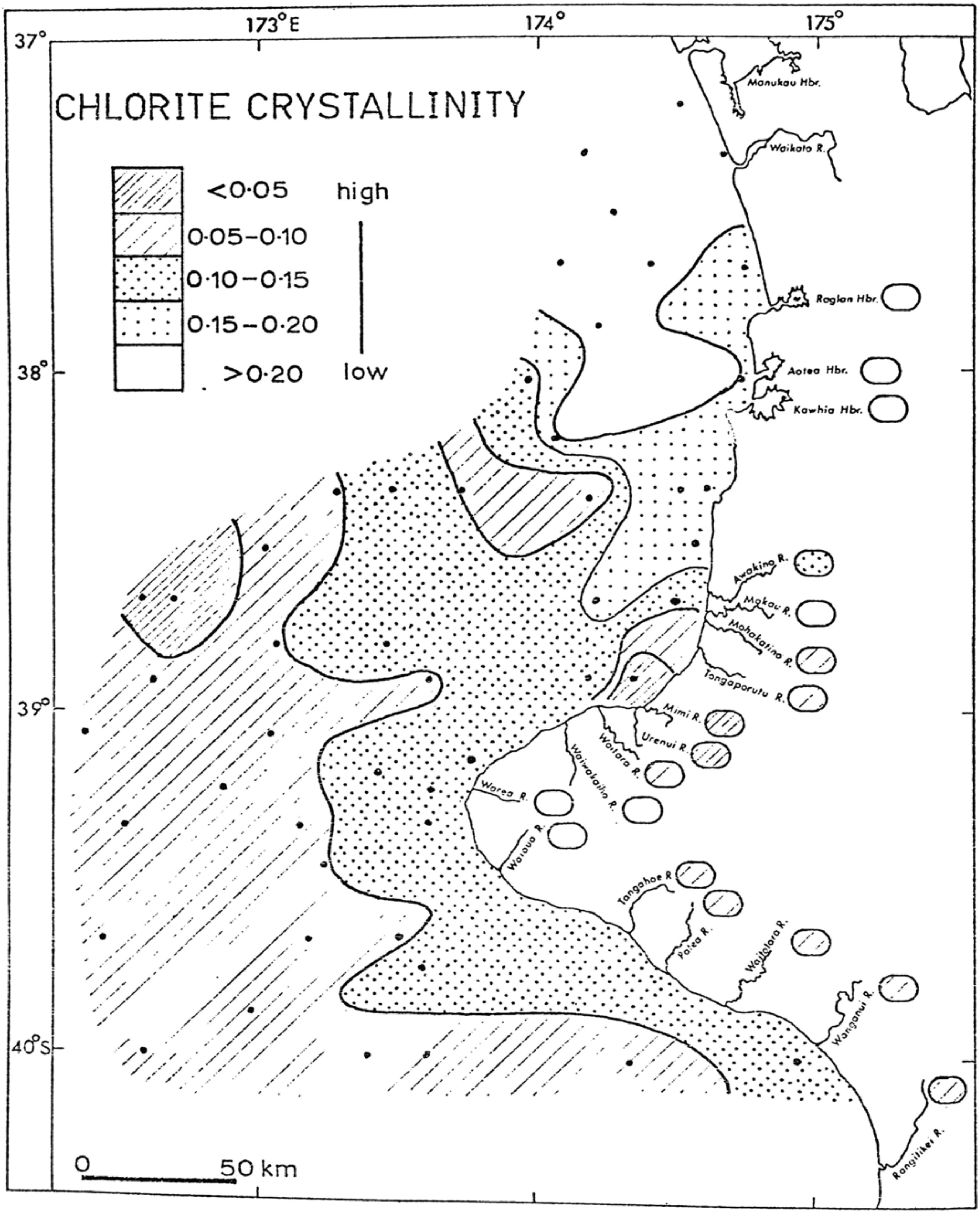


Fig. 4.13. Crystallinity of chlorite in $<2\mu$ size fraction of surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

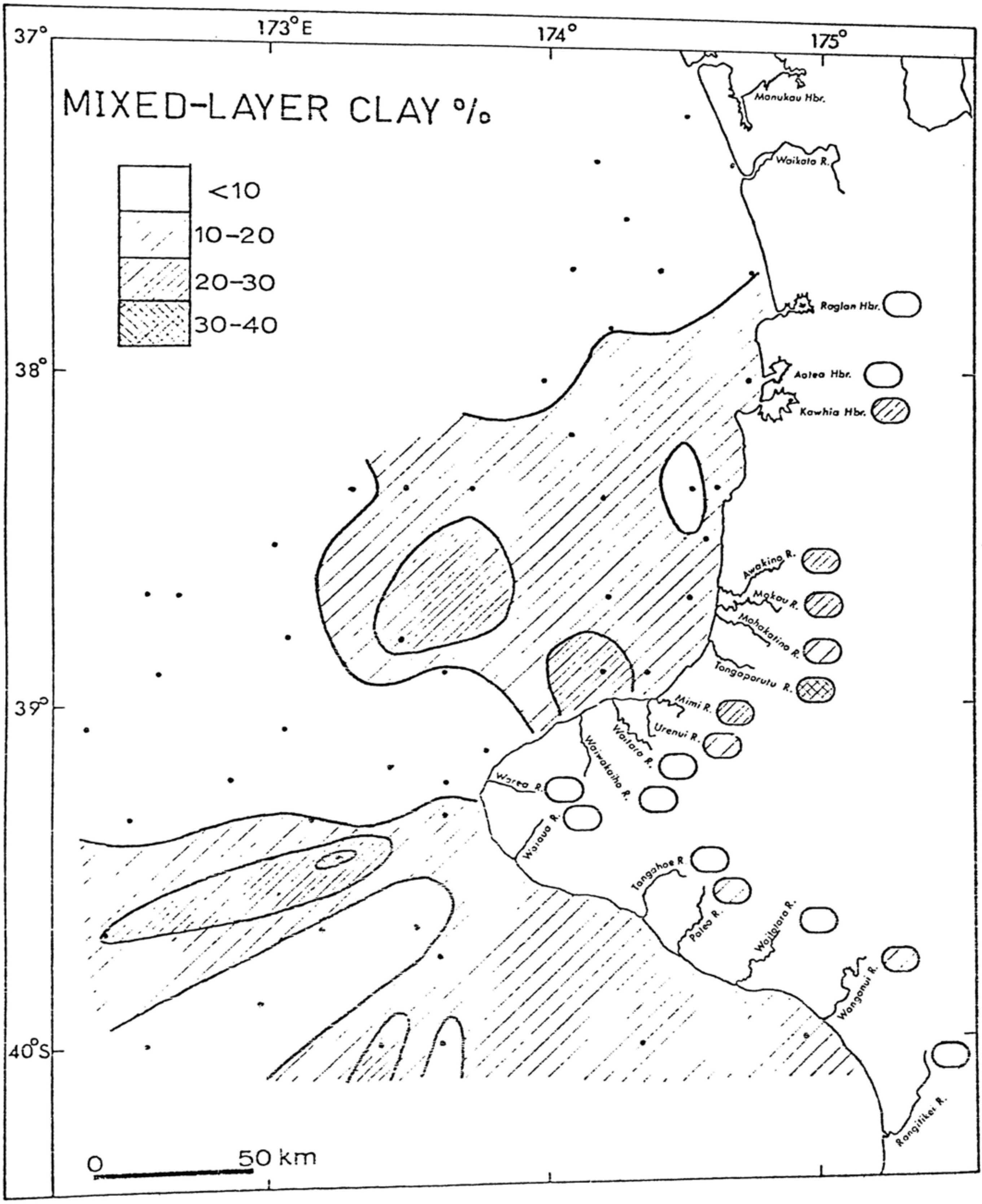


Fig. 4.14. Distribution of mixed-layer clay minerals (percentage of < 2µ size fractions) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

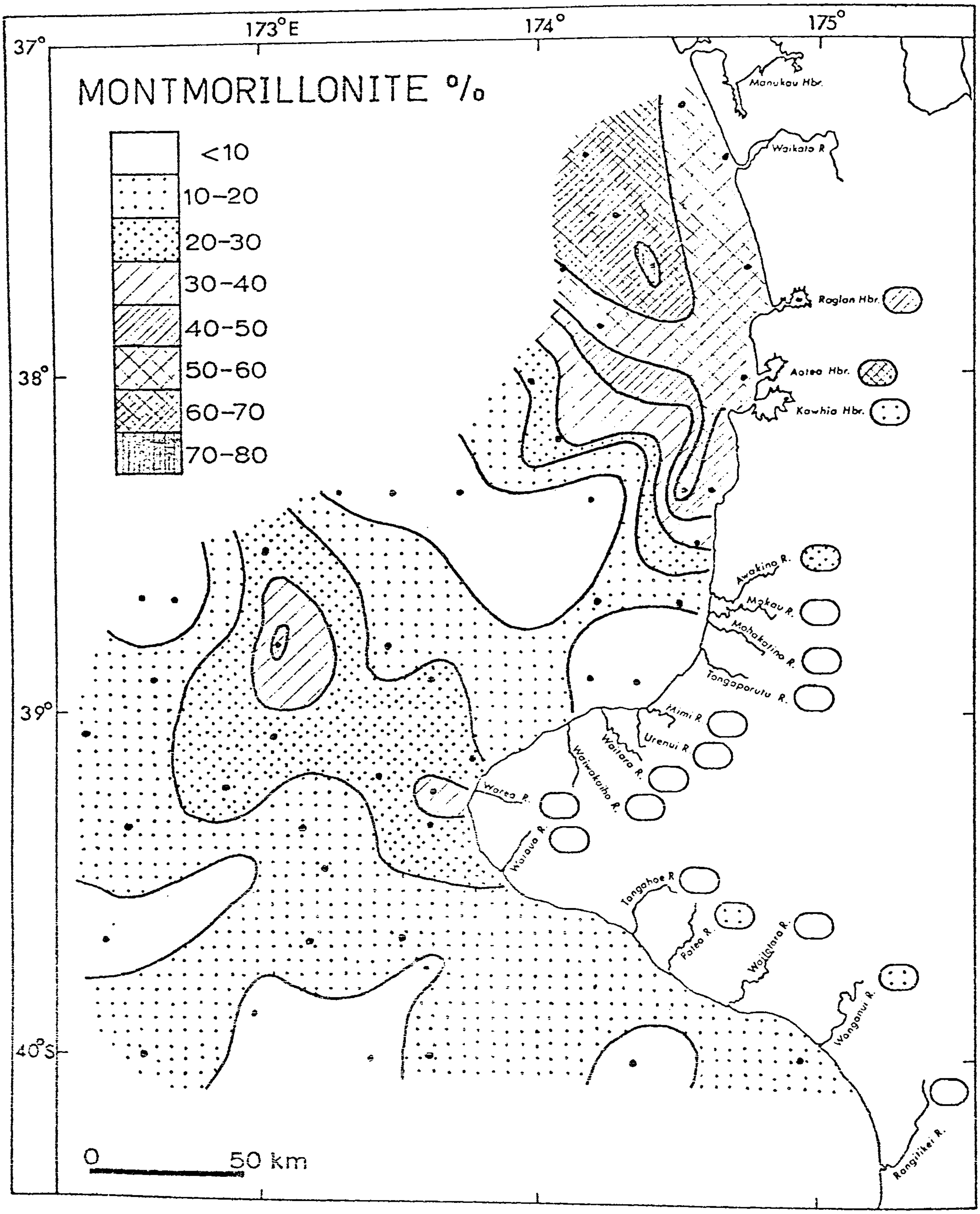


Fig. 4.15. Distribution of montmorillonite (percentage of $<2\mu$ size fractions) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

harbours and rivers. On the Hamilton Shelf, montmorillonite is notably more abundant at middle and outer shelf locations. Montmorillonite crystallinity shows an irregular distribution pattern

4.3 SEDIMENT DISTRIBUTION AND DISCUSSION

This section discusses the distribution and origin of sediments on the Hamilton, North Taranaki, Central Taranaki and South Taranaki Shelves and the North Cook Strait Basin. Sediment distribution patterns are explained mainly on the basis of sediment compositional data determined in this study and by comparing clay mineral distribution to present day seston dispersal patterns using satellite (LANDSAT) imagery. Analysis of LANDSAT imagery was made using the black and white band 4 spectral images and multi-band colour composite prints (1:1,000,000 scale).

4.3.1 HAMILTON SHELF SEDIMENTS

Hamilton Shelf sediments (Fig. 4.1) possess a characteristic clay mineral assemblage consisting of abundant to very abundant montmorillonite (40-80%), rare to very common poorly crystalline illite (0-30%), scarce to common kaolinite (5-20%), rare to scarce poorly crystalline chlorite (<10%) and rare to scarce mixed-layer clays (<10%) (Figs. 4.9 - 4.15; Pl. 4.2). This clay mineral suite is similar to that in the clay size fraction of sediments from the harbours bordering the Hamilton Shelf, suggesting a detrital origin for the shelf clay mineral assemblage. The anhedral habit of illite and chlorite supports a detrital origin for these minerals (Pl. 4.2). The harbours receive sediment from the erosion of Mesozoic sediments, Oligocene limestones, calcareous sandstones and calcareous mudstones plus Quaternary volcaniclastic sediments (Fig. 1.3 and Table 1.1). The main source of clays for the Hamilton Shelf sediments is probably the readily erodible Oligocene mudstones (Pls. 1.4 and 4.3) outcropping extensively on the coast and in the surrounding catchment (Fig. 1.3). The clay fraction of the widespread Whaingaroa Siltstone, for sample, contains 80-90% montmorillonite and some poorly crystalline illite and chlorite (Figs. 3.5 and 3.6; Pl. 3.13). The small amounts of kaolinite in Hamilton Shelf sediments are probably derived from the soils developed on Mesozoic sediments (see pp.105-107) and to a lesser extent from erosion of Waikato Coal Measures. The importance of the four large harbours (i.e. Manukau, Raglan, Aotea and Kauhia) to the shelf clay mineral budget can be assessed from LANDSAT imagery which shows suspended sediment plumes extending up to 20 km offshore from the mouths of these harbours. The plumes either extend straight out onto the shelf or show a significant up-coast drift (Pls. 4.4 - 4.8). The available imagery (30 October 1975; 15 and 16 February 1976) suggests that the prevailing nearshore surface currents on the Hamilton Shelf flow northwards - results compatible with those obtained by drift card measurements in the area (Brodie, 1960). Analysis of seston from shelf locations west of Raglan Harbour (Fig. 2.2; Samples N377-379) shows that terrigenous material is more abundant in nearshore samples. The nearshore seston (Samples N377 and 378) contains very common diatom fragments and clay material, chiefly montmorillonite, and rare glass shards (Pls. 4.9 -

4.12). The outer shelf seston (N379) contains unidentified amorphous material, rare detrital quartz and only minor amounts of clay suggesting that little terrigenous material reaches the outer shelf (Pl. 4.13).

The notably high concentrations of montmorillonite in Hamilton Shelf as compared to adjacent harbour sediments (Fig. 4.15) is possibly the result of differential flocculation and/or sedimentation of clay minerals on passing from the harbour to the shelf environment. A number of studies (e.g. Powers, 1957, Whitehouse et al., 1960; Porrenga, 1966; Edzwald and O'melia, 1975; Gibb, 1977) have attributed similar lateral changes in the offshore frequency distribution of clay minerals to these mechanisms. The physical chemistry of preferential flocculation and sedimentation was described elsewhere (pp. 117-119). In the case of the Hamilton Shelf montmorillonite evidence for such a process is meagre because: (1) the number of samples on which the distribution pattern is based is small (Fig. 4.15); (2) other clay minerals such as illite and chlorite show no systematic variations in abundance with distance offshore, a result contrary to which might be expected if preferential flocculation was an important process; (3) outer and middle shelf sediments are considered to be largely relict (McDougall and Brodie, 1967) implying that the montmorillonite distribution in these areas has no direct relationship to contemporary shelf processes; and (4) the large harbours, which are the major sediment source for the Hamilton Shelf, have salinities equal to or only slightly below those of adjacent shelf waters (Hume, 1977; Nelson, pers. comm. 1977), observations that suggest that any preferential flocculation would already have occurred before the clay minerals entered the shelf environment.

The possibility exists that the montmorillonite in the clay fraction of middle and outer shelf sediments is authigenic, being derived from the alteration of volcanic material (cf.pp.126-127) emplaced at these locations during Quaternary lower stands of sealevel. During the Quaternary, active erosion of the extensive rhyolitic and andesitic volcanic deposits of the Central North Island (Fig. 1.3; Table 1.1) provided considerable quantities of volcaniclastic material to large rivers such as the Waikato, which debouched onto the shelf (Hume et al., 1975). Winds also may have carried tephra material to the continental shelf and beyond. Over much of the inner shelf one would expect this material to be now buried by contemporary (Recent) sediments. At middle and outer shelf locations, where sediments are largely relict, any authigenic montmorillonite of volcanic origin may still be exposed and form part of the clay fraction along with detrital montmorillonite derived from the erosion of hinterland sediments, especially Oligocene mudstones. The origin of montmorillonite in Hamilton Shelf sediments is elaborated upon later in this chapter.

4.3.2 NORTH TARANAKI SHELF SEDIMENTS

North Taranaki Shelf sediments are characterised by a clay mineral assemblage containing abundant moderately crystalline illite (40-60%), common to very common mixed-layer clay minerals (10-30%), common moderately crystalline chlorite (10-20%), rare to common montmorillonite (0-20%)

and rare to scarce kaolinite (5-10%) (Figs. 4.9 - 4.15; Pl. 4.14). Figs. 4.9 - 4.15 show that the abundance and crystallinity of clay minerals in the North Taranaki Shelf sediments are comparable to those in the bordering river sediments (i.e. Awakino to Waitara Rivers) suggesting that the shelf clay mineral distribution patterns are related to regional changes in the onshore geology. Hinterland sediments are dominated by noncalcareous Miocene sandstones and mudstones (Fig. 1.3 and Table 1.1) whose clay minerals are chiefly illite, chlorite and mixed-layer clays (Figs. 3.5 and 3.6). In the north specific source rocks include mainly the Mahoenui mudstones (Pls. 1.9, 1.10 and 4.15), which account for the increasing amounts of montmorillonite in the northern North Taranaki Shelf sediments. In the south source rocks are the Mokau (Pl. 1.13), Mohakatino (Pls. 1.15, 1.16 and 1.18), Mt Messenger and Urenui (Pl. 1.21) mudstones, which produce the crystalline illite- and chlorite-rich clay fraction. Rapid erosion of the easily weathered mudstones is facilitated by the rugged topography (relief up to 600 m) and high rainfall (150 - 500 cm/yr) and erosional detritus is actively carried to the shelf before weathering significantly alters the clay mineralogy. A detrital origin for the clays is suggested by the predominance of anhedral illite and chlorite in the shelf sediments (Pl. 4.14). The smaller proportion of mixed-layer clays in North Taranaki Shelf sediments compared to adjacent river sediments (Fig. 4.14) may result from the transformation of detrital chlorite-vermiculite and chlorite-montmorillonite clays to their more stable chloritic analogues.

LANDSAT imagery of the area indicates that sediment being discharged from the rivers bordering the shelf produces suspended sediment plumes extending out some 40 km onto the shelf (Pls. 4.7 and 4.8); the shape of the plumes suggests there is a general southerly drift of suspended sediment on the North Taranaki Shelf. This is contrary to what might be expected if the major Westland Current (Fig. 4.2) exerted a significant influence on the direction of suspended sediment movement.

4.3.3 CENTRAL TARANAKI SHELF SEDIMENTS

Central Taranaki Shelf sediments are characterised by a clay fraction that contains abundant crystalline illite, (40-60%) very common montmorillonite (20-40%), common chlorite (10-20%), and rare to scarce mixed-layer clays (< 10%) and rare kaolinite (< 5%) (Figs. 4.9 - 4.15; Pls. 4.16 and 4.17). The clay fraction of sediments from rivers that flank the Central Taranaki Shelf (i.e. the Waiwakaiho to Waiaua Rivers) contain no crystalline clay minerals (Appendix Table IV.1; Fieldes, 1968), suggesting that the clays of the shelf are unlikely to be inherited from these rivers and their surrounding catchments. Rivers flanking the Central Taranaki Shelf drain mainly Quaternary andesitic volcaniclastic sediments whose soils have a clay mineralogy dominated by allophane (Symes and Wells, 1973).

Possible origins for the shelf clay mineral suite include: (1) derivation from rivers entering the North and South Taranaki Bights (Fig. 4.1), with subsequent transportation of the clays onto the shelf area by ocean currents;

(2) derivation from a northern South Island source; (3) reworking from adjacent North and South Taranaki Shelf areas; and (4) authigenisis.

Consideration of the clay mineral suites being discharged from rivers of the North and South Taranaki Bights (Figs. 4.9 - 4.15) indicates that they are a possible source for the crystalline illite and chlorite, montmorillonite and kaolinite in the Central Taranaki Shelf sediments. If the major circulation patterns on the shelf (Fig. 4. 2) are important in controlling the directions of sediment movement then it would appear that clay material emerging from the North and South Taranaki Bight rivers is unlikely to be swept towards the Central Taranaki Shelf. However, when strong southerly winds minimise the indraught of the D'Urville Current into Cook Strait clay material could be transported northwards from the South Taranaki Bight rivers to the Central Taranaki Shelf. Moreover, in the case of the North Taranaki Bight, LANDSAT imagery (Pls. 4.7 and 4.8) shows that suspended sediment plumes from the Awakino to Tongaporutu Rivers (Fig. 4.1) extend up to 40 km across the shelf and have a southerly drift component opposite in direction to the offshore Westland Current (c f. Fig. 4.2). Presumably, under favourable conditions, some of this material could ultimately be deposited on the Central Taranaki Shelf. The predominance of anhedral particles of illite and chlorite in the Central Taranaki Shelf sediments is compatible with a detrital origin (Pls. 4.16 and 4.17).

It is, however, unlikely that the montmorillonite in the clay fraction of Central Taranaki Shelf sediments is entirely detrital. Adjacent source areas are by comparison relatively depleted in montmorillonite (Fig. 4.15). An alternative source for the montmorillonite is from the marine alteration of volcanogenic materials. The Mount Egmont region consists entirely of volcanogenic deposits (Fig. 1.3) and soils are rich in allophanic clays and glass shards (pp.10-12). Rivers and streams discharge this material directly onto the shelf where it may alter to montmorillonite. Surprisingly, allophane appears to be absent from the clay size fraction of Central Taranaki Shelf sediments. However samples do show an abundance of fine grained, lath-shaped flakes of montmorillonite and some possible amorphous siliceous material (Pls. 4.16 and 4.17). Similar features were shown earlier to be characteristically associated with neoformed montmorillonite (cf. Pls. 4.16 and 4.17 with Pl. 3.21 and Table 3.6). Central Taranaki Shelf sediments contain significant amounts of X-ray amorphous material in nearshore locations (Fig. 4.8). The occurrence of titanomagnetite, hornblende and andesitic rock fragments in these sediments strongly suggests the amorphous material is of volcanic origin. To assess in more detail the nature of material reaching the shelf three seston samples from the nearshore zone were examined (Pls. 4.18 and 4.19); samples were not available from middle and outer shelf locations. The seston contains abundant diatom fragments and organic matter, common crystalline clay minerals (usually in aggregates or floccules), but apparently no allophane or glass shards. Glassy volcanic material readily alters to montmorillonite in the marine environment and in fact may alter during transport (Dunoyer de Segonzac, 1970, p.293; cf. pp.173-174). The absence of allophane in the clay fraction of the Central Taranaki Shelf sediments and its surficial abundance in adjacent onshore areas raises the question of its fate on entering the marine environment. It is possible that allophane alters readily to montmorillonite in the same manner as glass shards do (cf. pp.173-174). The origin of montmorillonite is discussed further in section 4.3.6.

Montmorillonite in the Central Taranaki Shelf sediments is thus considered to be both detrital and neoformed.

4.3.4 SOUTH TARANAKI SHELF SEDIMENTS

The clay fraction of South Taranaki Shelf sediments contains abundant to very abundant crystalline illite (50-60%), very common mixed-layer clays (20-40%) and crystalline chlorite (20-40%), rare to common montmorillonite (0-20%) and rare kaolinite (< 5%) (Figs. 4.9 - 4.15; Pl. 4.20). Because of its location (Fig. 4.1) it is unlikely that the South Taranaki Shelf receives much terrigenous sediment directly from the western North Island rivers. More likely sources of clays are the northwestern South Island, reworking of adjacent Central Taranaki Shelf and North Cook Strait Basin sediments and, perhaps, rivers of the South Taranaki Bight. Unfortunately there is little data available on the nature of clay minerals in northwestern South Island rocks, which comprise a mixture of Lower Paleozoic sandstones and mudstones and schists, Middle Paleozoic chlorite and mica schists, Upper Paleozoic granites and grandiorites and Cenozoic sediments (Grindley, 1961).

Analysis of LANDSAT imagery (Pls. 4.21 - 4.23) shows that suspended sediment is carried from the western and northwestern coasts of the South Island to the South Taranaki Shelf and Cook Strait Basin, presumably by the Westland and D'Urville Currents (Fig. 4.2). No evidence exists for significant suspended sediment contributions to the South Taranaki Shelf from rivers of the South Taranaki Bight.

While there is little data available on the clay mineralogy of sediments from the northern and western South Island it is likely that crystalline illite and chlorite dominate. Bulk chemical analyses show that the K_2O content of western shelf sediments is highest in the southwest (Appendix Fig. VIII.4), presumably partly due to the large quantity of crystalline illite in the clay size fraction of sediments (Figs. 4.10 and 4.11). Optical examination of the sand - and silt-size fractions of the sediments shows appreciable quantities of micas that were probably derived from the erosion of plutonic rocks in north-western South Island and subsequently carried north by the Westland Current. The K_2O distribution in shelf sediments indicates that micas are perhaps carried as far north as Cape Egmont. Chlorite is most abundant and most crystalline in the clay fraction of sediments in the southwest of the study area (Figs. 4.12 and 4.12). This probably results from the Westland Current transporting chlorite northwards from a northwestern South Island source. A detrital origin for the illite and chlorite in South Taranaki Shelf sediments is supported by the anhedral habit of these clays (Pl. 4.20).

The Central Mud Zone of the South Taranaki Shelf (Fig. 4.3) is a zone of comparatively quiet water produced by a pronounced change in the pattern of tidal and ocean currents in the area (Lewis and Eade, 1974). Clay mineral and other data from this study indicate that suspended sediment from the northwestern South Island is either transported eastwards by the D'Urville Current to the North Cook Strait Basin, and perhaps on to Cook Strait, or transported northeastwards by the Westland Current to the Mud Zone where it is deposited. The distribution pattern of clays (Figs. 4.10-4.13) and K_2O (Appendix Fig. V111.4) on the South Taranaki Shelf indicates that mud is preferentially deposited in the southern portion of the present Central Mud Belt. Some illite and chlorite is transported further to the northeast, particularly when strong southerly winds minimise the indraught of the D'Urville Current into Cook Strait, where they are deposited on the Central Taranaki Shelf contemporaneously with authigenic montmorillonite in a zone of relatively low terrigenous sedimentation.

4.3.5 NORTH COOK STRAIT BASIN SEDIMENTS

The clay mineralogy of North Cook Strait Basin sediments consist of abundant to very abundant crystalline illite (50-60%), common mixed-layer clay minerals (10-20%) crystalline chlorite (10-20%) and montmorillonite (10-20%), and rare kaolinite ($< 5\%$) (Figs. 4.9 - 4.15). Figs. 4.9 - 4.15 show that sediments from the North Cook Strait Basin contain clay minerals in similar abundance and with similar crystallinity to those in adjacent river sediments (i.e. Tangahoe to Rangitikei Rivers) suggesting that the clay mineral distribution patterns are provenance controlled (Fig. 1.3 and Table 1.1). Most rivers flanking the North Cook Strait Basin drain catchments of Pliocene - Pleistocene sediments, although the major Wanganui River cuts back into older Mesozoic to Miocene sediments. Onshore sediments are mainly easily erodible mudstones of the Mt Messenger (Pl. 1.20), Urenui (Pl. 1.21) and Tangahoe (Pls. 1.23 and 4.24) formations and Lower Quaternary deposits (Fig. 1.3). The clay mineralogy of these sediments is dominated by crystalline illite with very common montmorillonite and common crystalline chlorite, scarce mixed-layer clays and rare kaolinite (Figs. 3.5 and 3.6), features similar to adjacent shelf assemblages. Rivers in the catchment areas are deeply entrenched (Fig. 3.2) in Miocene and Pliocene sediments exposing tens of metres of frittering mudstones capped by thin deposits of Quaternary alluvium, tephra or soil. Inland, relief up to 600 m and the moderately high rainfall (125 - 250 cm/yr) ensure rapid erosion of the mudstone slopes so that there is minimal alteration of the clay minerals transported to the Cook Strait Basin. Other studies (McDougall and Brodie, 1967; Carter, 1975) have suggested that in the North Cook Strait Basin the input of modern terrigenous mud from rivers of the South Taranaki Bight is so high that the hydrodynamic regime cannot adequately disperse all the sediment with the result that mud blankets relict terrigenous and biogenic deposits.

LANDSAT imagery indicates that suspended sediment being discharged from the rivers bordering the North Cook Strait Basin forms plumes extending about 10 km offshore

(Pls. 4.25 and 4.26). North of the Wanganui River plumes extend directly out from the coast while south of the River suspended sediment clearly moves southwards along the coast (Pl. 4.25) suggesting the influence of a northerly branch of the D'Urville Current (Fig. 4.2). Both textural studies (McDougall and Brodie, 1967; Lewis and Eade, 1974) and LANDSAT imagery suggest that at least some of this sediment is swept south to Cook Strait (Pl. 4.25). Available LANDSAT imagery indicates that most sediment from South Taranaki Bight Rivers is deposited relatively close to the coast and not in the Central Mud Zone of the Cook Strait Basin as suggested by Carter (1975). This implies that much of the clay material in the North Cook Strait Basin ultimately has a northwestern South Island provenance.

4.4 FACTORS CONTROLLING CLAY MINERAL DISTRIBUTIONS

It appears that the frequency distribution patterns of clay minerals in western shelf sediments (Figs. 4.9 - 4.15) can be largely interpreted in terms of changes in the clay mineralogy of adjacent onshore sediment sources. For example the importance of an onshore provenance for kaolinite is unequivocal, for while kaolinite is stable in the marine environment it is unlikely to form there (Keller, 1970). The increased abundance of kaolinite in northern western shelf sediments is paralleled by a similar increase in kaolinite in the clay fraction of river and harbour sediments. However a simple provenance control on the clay mineral distributions is complicated by the occurrence of both modern and relict sediments on the western shelf (McDougall and Brodie, 1967), by the presence of authigenic materials, and by the locally variable hydrologic regime. The following discussion considers the relative roles that provenance and shelf processes play in controlling the present frequency distribution patterns of clay minerals on the western shelf.

4.4.1 PROVENANCE

The frequency distribution of clay minerals in shelf sediments is dependent in part on the type and availability of clay material in onshore catchment areas.

Comparatively little clay material is thought to be detritally inherited from soil and tephric units because XRD, TEM and SEM analyses show that the shelf clay mineralogy is more comparable to that of the onshore sediments. The absence of distinct volcanic beds in western shelf sediments and the small proportion of glass shards in the sediments indicate that volcanism was not a significant source in terms of quantity, that volcanic materials have been dispersed by the hydraulic regime and diluted with other sediments by shelf processes, and/or that the glassy and finest ash material has been altered to other minerals.

It is likely that mudstones supply the bulk of the clay material to western North Island rivers and harbours, and ultimately to the shelf, because they are the most abundant lithology onshore, they contain significantly more clay minerals than other lithologies, and they are generally readily erodible. Clay mineralogic data suggest that Hamilton Shelf clays are derived largely from Oligocene mudstones, and North Taranaki Shelf clays from Miocene mudstones. North Cook Strait Basin clays, on the other hand, are supplied partly from Upper Miocene-Pliocene-Quaternary mudstones and partly from western South Island derived seston, while South Taranaki Shelf clays are probably also supplied mainly from the latter source.

4.4.2 RELICT VERSUS RECENT SEDIMENTS

McDougall and Brodie (1967), Lewis and Eade (1974) and Carter (1975) have shown that modern terrigenous sediments north of Cape Egmont are represented by a coastal zone of very fine sands and muds (Fig. 4.3) extending some 10-30 km onto the shelf. The outer shelf very fine sand zone and the middle shelf fine sand zone are, on the basis of their texture, presumed to be relict sediments deposited during and/or following the Last Glacial maximum (McDougall and Brodie, 1967). The carbonate-rich deposits west of the Mokau River mouth are formed from abundant relict skeletal material consisting mainly of coarse molluscan and bryozoan fragments. On the outer Hamilton Shelf the carbonate material is dominated by coccolith tests (Pl. 4.1; McDougall, 1972). The fact that little terrigenous material is presently being supplied to the outer Hamilton Shelf is indicated by seston analyses (Pls. 4.9 - 4.13).

Clay mineral distribution patterns (Fig. 4.7) bear little relationship to the distribution of muddy sediments (Fig. 4.3). Textural studies (McDougall and Brodie, 1967) indicate that mud is mainly deposited close to shore as a result of flocculation of clay material emerging from river mouths onto the shelf. LANDSAT imagery north of Cape Egmont supports this contention, in as far as it shows suspended sediment emerging from the mouths of harbours and rivers and extending only some 10-40 km out onto the shelf (Pls. 4.4 -

4.8). Significantly the plumes extending out from the harbours bordering the Hamilton Shelf may drift north at the same time as those from rivers bordering the North Taranaki Shelf drift south (Pls. 4.7 and 4.8). This suggests that over the inner and perhaps middle shelf, surface current directions are influenced little or not at all by major current systems (cf. Fig. 4.2). The southerly directed currents may result from local interactions of ocean, tide and wind generated currents, or in the case of North Taranaki Shelf, from a possible southerly directed backwater effect produced by the northerly passage of the major Westland Current past Cape Egmont. Satellite imagery has revealed a similar effect on the eastern shelf of the South Island where the northerly directed Southland Current flows past Banks Peninsula (A.J.R. Male pers. comm. 1977).

The featureless morphology of the western shelf surface probably exerts little influence on the distribution of clay minerals. Northwards from Cape Egmont, clay mineral abundance and crystallinity zones on the shelf correspond closely with those of adjacent onshore provenance areas, suggesting that bottom currents have little effect on distributing clays in middle and outer shelf positions. Deviations from this simple pattern occur nearshore where the effects of river run-off, tides and winds produce a nett movement and resultant mixing of material. The extent of this mixing is indicated by LANDSAT imagery of suspended sediment plumes emerging from harbour and river mouths. The writer considers that on the western shelf north of Cape Egmont, clay minerals on the outer and most likely the middle shelf areas are probably relict. In these regions the clay minerals were probably deposited during lower sealevel stands, or immediately following the Last Glaciation (Otiran Stage in New Zealand). At that time rivers drained similar source lithologies to the present and subaerially exposed inner shelf areas now covered by modern sediments. The rivers debouched their sediment loads onto what was then the nearshore zone. The subsequent rise in sealevel (Aranuian Stage in New Zealand) has left these sediments as essentially relict middle and outer shelf deposits. As the same source lithologies existed throughout the Quaternary, there are no significant changes between inner, middle, and outer shelf clay mineral assemblages.

In summary, north of Cape Egmont the clay materials on the middle and outer shelf are probably mainly relict while those on the inner shelf are Recent.

To the west and south of Cape Egmont the shelf sediments include an outer zone of muddy very fine sands, an extensive zone of sandy muds and muds and, over much of the southern part of the shelf, muddy fine sands, together with a prominent area of coarse sands and gravels in the South Taranaki Bight (Fig. 4.3). The outer very fine sand zone is considered to represent relict material (McDougall and Brodie, 1967). As in the case of the outer shelf sediments north of Cape Egmont, these sediments have a high carbonate content, comprising abundant foraminiferal and coccolith material, indicating the area experiences low rates of terrigenous sedimentation. Reduced terrigenous sedimentation apparently also characterises the area of coarse-grained sediments in the North Cook Strait Basin which consists largely of both relict and Recent skeletal material (McDougall and Brodie, 1967; Lewis and Eade, 1974). This area coincides with the major eastward flowing branch of the D'Urville Current (Fig. 4.2) which serves to keep the sediment relatively free from Recent fine terrigenous material. It is possible therefore that some of the clay material in this area is also relict. In the nearshore region the North Cook Strait Basin clay material is undoubtedly modern because LANDSAT imagery shows sediment emerging up to 10 km onto the shelf (Pls. 4.25 and 4.26). LANDSAT imagery (Pls. 4.22 and 4.23) and clay mineralogic data indicate that clays in the Mud Zone on the South Taranaki shelf are also modern, being derived principally from rivers draining the west and northwest coast of the South Island with material subsequently being carried north-east by the Westland Current. In the offshore parts of the North Cook Strait modern clays of similar composition to those on the South Taranaki shelf are likewise deposited from the offshoot D'Urville Current.

Comparison of the total clay mineral distribution pattern (Fig. 4.7) with the distribution patterns of individual clay minerals (Figs. 4.9 - 4.15) indicates that the two areas of concentration of clay minerals in the Mud Zone west and southwest of Cape Egmont (cf. Figs. 4.3 and 4.7) are characterised by montmorillonite-rich and montmorillonite-poor material respectively. Clay mineralogic evidence suggests that the southern montmorillonite-poor area results from the influx of crystalline illite and chlorite from a western South Island provenance. The northern montmorillonite-rich zone on the Central Taranaki Shelf similarly receives illite and chlorite from the same source but these clays are diluted by locally derived authigenic montmorillonite.

4.4.3 AUTHIGENESIS

The possible contribution of authigenesis to the present distribution pattern of western shelf clays has been mentioned earlier (pp.217-225). TEM data for western shelf clays generally suggest that illite and chlorite are detrital in origin. Kaolinite is most probably derived from an onshore source as it is unlikely to form in the marine environment (Keller, 1970). Certain mixed-layer clay mineral species may be transformed in the marine environment (cf. p.221). It is considered that montmorillonite in western shelf sediments is of both detrital and authigenic origin, with neoformed volcanogenic montmorillonite being concentrated in sediments of the Central Taranaki and Hamilton Shelves (Fig. 4.15).

The montmorillonite component of all samples was examined in more detail in an attempt to determine the factors controlling the distribution of detrital and authigenic montmorillonite. Many western shelf sediments contain both dioctahedral ("non-volcanic") and trioctahedral ("volcanic") montmorillonite. A number of samples contain lath-shaped flakes of montmorillonite and these are especially common in Central Taranaki Shelf sediments (Pls. 4.16 and 4.17). The crystallinity of montmorillonite on the shelf varies, but the most highly crystalline material occurs in the zones of relatively high montmorillonite concentration off Raglan Harbour and Cape Egmont. Plagioclase species A and B are present in many samples, a feature characteristic of onshore Quaternary volcanogenic sediments (p.78), but plagioclase A is usually most abundant. The generally random distribution on the shelf of all these properties, which otherwise proved useful in characterising neoformed montmorillonite in onshore samples, would suggest that authigenic montmorillonite occurs at many localities on the shelf. The absolute concentration of authigenic montmorillonite in the clay fraction is therefore governed primarily by dilution by detrital clay minerals, including detrital montmorillonite. The relative abundance of montmorillonite in middle and outer Hamilton Shelf sediments, and particularly in Central Taranaki Shelf sediments (Fig. 4.15), is considered to reflect concentration of authigenic montmorillonite in areas where dilution by detrital clays is low.

The clay mineralogy of two 1 m cores of mud from the Central Taranaki Shelf (localities C422 and C429; Fig. 2.2) varies little (cf. Appendix Table IV.1 and IV.2) from top (C422.1 and C429.1) to bottom (C422.4 and C429.4) and is dominated by illite with common chlorite and smaller amounts of montmorillonite, mixed-layer clays and kaolinite. The cores show little evidence of compaction and presumably represent only a short time interval. If the montmorillonite in these samples is authigenic then its similarity in abundance down the core implies montmorillonite formation was early diagenetic (cf. Dunoyer de Segonzac, 1970, p.293).

4.4.4 REWORKING

From the available data it is difficult to assess the extent to which reworking has affected the distribution of clay minerals in western shelf sediments. Reworking of sediments could be invoked to explain any irregularities in the patterns of clay mineral distributions. At middle and outer shelf locations in areas of relict sediments, much of the existing clay material may be protected from erosion by a lag-cover of coarse sediment. Boundary layer turbulence, associated with coarse-textured relict deposits and high current velocities resulting from tide and current interaction, tend to prevent deposition of mud at middle and outer shelf locations (Van der Linden, 1969). Sediment movement in these areas could take place under the present hydrodynamic regime as grounding storm swells on the Taranaki Shelf may resuspend bottom material to depths of at least 130 m during the 25 year storm event (Glenn et al., 1973). The resuspended material can then be moved by bottom currents. However, if reworking was an important process in controlling the distribution of clay minerals in western shelf sediments, one might have expected more uniformly mixed clay mineral assemblages on the shelf.

4.5 SUMMARY AND CONCLUSIONS

The western shelf is divided into five zones: the Hamilton Shelf, the North Taranaki Shelf, the Central Taranaki Shelf, the South Taranaki Shelf and the North Cook Strait Basin (Fig. 4.1).

Western shelf surficial sediments are mainly muddy fine sands, muddy very fine sands and sandy muds (Fig. 4.3). The Hamilton and North Taranaki Shelf sediments are dominantly modern terrigenous sands and muds on the inner shelf, relict biogenic and terrigenous sands on the middle shelf and mixed modern/relict biogenic sands and muds on the outer shelf. Mixed modern/relict mud occupies most of the Central and South Taranaki Shelves. North Cook Strait Basin sediments are dominantly mixed modern/relict biogenic and terrigenous sands and gravels.

The $CaCO_3$ fraction of western shelf sediments consists almost exclusively of biogenic material that forms significant concentrations (20-30%) at certain middle shelf (areas of relict skeletal debris) and outer shelf (areas of low terrigenous input) locations (Fig. 4.4). Quartz, plagioclase feldspar, and clay minerals are the most abundant constituents in the non-biogenic fraction.

The clay fraction of western shelf sediments consists mostly of illite and montmorillonite, lesser amounts of chlorite and mixed-layer clays, and some kaolinite. In general the crystallinity of both illite and chlorite increases towards the south.

The distribution and origin of clay minerals established in this study supports the recognition of five petrologic zones on the western shelf (Fig. 4.1):

(1) Hamilton Shelf - Abundant montmorillonite, rare to common poorly crystalline illite and lesser kaolinite, mixed-layer clays and poorly crystalline chlorite. Montmorillonite is primarily detrital from Oligocene mudstones with some montmorillonite perhaps being neoformed from Quaternary volcanogenic material.

(2) North Taranaki Shelf - Abundant crystalline illite, common mixed-layer clays and crystalline chlorite, minor montmorillonite; primarily detrital from Miocene mudstones.

(3) Central Taranaki Shelf - Abundant crystalline illite, very common montmorillonite, common chlorite and lesser mixed-layer clays and kaolinite. Neoformation of montmorillonite from Egmont - derived volcanogenic material contemporaneous with deposition of detrital illite, montmorillonite, chlorite and mixed-layer clays. The detrital clays are derived mainly from northwestern South Island rocks and sediments via the Westland Current and to a lesser extent from a North and South Taranaki Bight terrestial provenance.

(4) South Taranaki Shelf - Abundant crystalline illite and common mixed-layer clays and crystalline chlorite; primarily detrital from northwestern South Island rocks and sediments and carried north by the Westland Current.

(5) North Cook Strait Basin - Abundant crystalline illite, common mixed-layer clays and crystalline chlorite and montmorillonite; primarily detrital from northwestern South Island rocks and sediments via the D'Urville current and to a lesser extent from Pliocene-Quaternary mudstones in the south Taranaki - Wanganui district.

In Hamilton and North Taranaki Shelf sediments, clay mineral distribution patterns reflect regional changes in onshore geology and the clays are largely detrital. The outer and possibly the middle shelves are areas of dominantly relict clay minerals while inner shelf areas represent zones of contemporary clay mineral deposition. In Central Taranaki Shelf, South Taranaki Shelf and North Cook Strait Basin sediments the clays have a more complex origin. The northwestern South Island supplies detrital clays to all three zones but on the Central Taranaki Shelf where the terrigenous input is low, montmorillonite is neoformed from volcanogenic material.

Analysis of the frequency distribution patterns of clay mineral species on the western shelf shows that there is no simple relationship between clay mineral distribution, clay mineral provenance or environmental conditions. This is because western shelf sediments are the product of the interplay of a variety of physical, chemical and biochemical processes that have produced both detrital and authigenic sediments. Also, shelf sediments represent both modern and relict material.

Sattelite imagery is considered to be a useful tool for determining clay mineral provenance and suspended sediment distribution patterns in Recent sediments. However, the data should be applied with an appreciation that bottom currents may differ from surface currents and, furthermore, because of the complex interplay of factors controlling clay mineral distributions in surficial sediments, sattelite imagery is best used in conjunction with standard petrologic techniques in studies of bottom sediments.

FUTURE WORK

The information obtained in this study suggests a number of lines of research that are worthy of future investigation.

1. Comparison of clay mineral data obtained in this study, particularly those aspects relating to the effects of source and environment on clay mineral distributions, with data from:

(a) The western shelf sediments at depth.

(b) Other parts of the New Zealand continental shelf, particularly those areas where different source or environmental conditions exist.

(c) The subsurface equivalents of the Cenozoic sediments exposed in the Taranaki and Wanganui Districts.

(d) Cenozoic sequences from other parts of New Zealand, particularly those areas having different tectonic histories such as in Southland and Northland where tectonism influenced sedimentation in the Oligocene, or along the East Coast of the North Island where sedimentation has been strongly influenced by both tectonism and volcanism.

2. The application of clay petrologic data obtained from the above mentioned studies to the fields of ceramics, and engineering and petroleum geology.

REFERENCES

ALMON, W.A.; FULLERTON, L.B.; DAVIES, D.K. 1976: Pore space reduction in Cretaceous sandstone through chemical precipitation of clay minerals. Journal of Sedimentary Petrology 46: 89-96.

AOYAGI, K. 1967: Mineralogical study of sedimentary rocks in the oil fields of Japan by the XRD method, and its application to petroleum geology. Clay Science 3 : 37-54.

BALLANCE, P.F. 1976: Review of "Present plate boundary and its evolution in the New Zealand Region". - Reply. New Zealand Journal of Geology and Geophysics 19: 755-8.

BARNHISEL, R.I.; RICH, C.I. 1967: Clay mineral formation in different rock types of a weathering boulder conglomerate. Proceedings of the American Soil Science Society 31: 627-631.

BARSHAD, I. 1964: Chemistry of soil development. Pp. 1-70 in Bear, F.E. (Ed.): Chemistry of the Soil Reinhold, New York.

BEAVAN, P.J.; DUMBLETON, M.J. 1966: Clay minerals and geomorphology in four Caribbean Islands. Clay Minerals 6 : 371-382.

BERGGREN, W.A. 1972: A Cenozoic time-scale some implications for regional geology and paleobiogeography. Lethia 5 : 195-215.

BEU, A.G.; HENDERSON, R.A.; NELSON, C.S. 1972: Notes on the taphonomy and paleoecology of New Zealand Tertiary spatangoida. New Zealand Journal of Geology and Geophysics 15: 275-286.

BEUTELSPACHER, H.; VAN DER MAREL, H.W. 1968: "Atlas of Electron Microscopy of Clay Minerals and Their Mixtures". Elsevier, Amsterdam. 332p.

BIRKELAND, P.W. 1969: Quaternary paleoclimatic implications of soil clay mineral distribution in a Sierra Nevada - Great Basin transect. Journal of Geology 77: 289-302.

BISCAYE, P.E. 1964: Distinction between kaolinite and chlorite in Recent sediments by X-ray diffraction. American Mineralogist 49: 1281-1289.

BISCAYE, P.E. 1965: Mineralogy and sedimentation of Recent deep-sea clay in the Atlantic Ocean and adjacent seas and oceans. Bulletin of the Geological Society of America 76 : 803-832.

BLATT, H.; MIDDLETON, G.; MURRAY, R. 1972: "Origin of Sedimentary Rocks." Prentice Hall, Inc., New Jersey. 634p.

BOUMA, A.H. 1962: Sedimentology of Some Flysch Deposits. Elsevier, Amsterdam. 168p.

BRADLEY, W.F. 1953: Analysis of mixed-layer clay mineral structures. *Analytical Chemistry 25* : 727-730.

BRINDLEY, G.W. 1967: Disorder in Phyllosilicates. *in* "Layer Silicates." A.G.I. Short Course Lecture Notes, American Geological Institute, Washington, D.C.

BRINDLEY, G.W. 1972: Chlorite minerals. Pp. 242-296 *in* Brown, G. (Ed.) : "X-ray Identification and Crystal Structures of Clay Minerals." Mineralogical Society, London. 544p.

BRODIE, J.W. 1960: Coastal surface currents around New Zealand. *New Zealand Journal of Geology and Geophysics 3* : 235-252.

BROWN, G. 1953: The dioctahedral analogue of vermiculite. *Bulletin of Clay Mineralogy 2* : 64-69.

BROWN, G. 1955: The effect of isomorphous substitutions on the intensities of (001) relections of mica- and chlorite-type structures. *Mineral Magazine 30* : 657-665.

BROWN, G. (Ed.) 1972: "The X-ray Identification and Crystal Structures of Clay Minerals." Mineralogical Society, London. 544p.

BRUNTON, G. 1955: Vapour pressure glycolation of oriented clay minerals. *American Mineralogist 40* : 124-126.

BURST, J.F. 1958: Mineral heterogeneity in "glauconite" pellets. *American Mineralogist 43* : 481-497.

BURST, J.F. 1959: Post diagenetic clay-mineral environmental relationships in the Gulf Coast Eocene. Pp. 327-341 *in* Swineford, A. (Ed.) : *Clays and Clay Minerals*, Proceedings of the 6th National Conference, Pergamon Press, New York.

BURST, J.F. 1969: Diagenesis of Gulf Coast clayey sediments and its possible relation to petroleum migration. *Bulletin of the American Association of Petroleum Geologists 53* : 73-93.

BYERS, C.W. 1974 : Shale fissility : relation to bioturbation. *Sedimentology 21* : 479-484.

CARLSON, J.R.; ROGERS, K.A. 1973: Some difficulties in the quantitative X-ray modal analysis of rocks by the method of Nelson and Cochrane (1970). *Tane 19* 221-222.

CARROLL, D. 1970: Clay Minerals : A Guide to their X-ray Identification. Geological Society of American Special Paper 126. 88p.

CARROLL, D.; STARKEY, H. 1960: Effect of sea water on clay minerals. Pp. 80-101 in Swineford, A. (Ed.): Clays and Clay Minerals, Pergamon Press, New York.

CARROLL, D.; STARKEY, H.C. 1971: Reactivity of clay minerals with acids and alkakies. Clays and Clay Minerals 19 : 321-335.

CARTER, L. 1975: Sedimentation on the continental terrace around New Zealand : a review. Marine Geology 19 : 209-238.

CARTER, R.M. 1974 : The moulding of the landscape (3). Pp. 211-217 in "New Zealand's Nature Heritage". Volume 2, Hamlyn House Books, New Zealand. 284p.

CARTER, R.M.; NORRIS, R.J. 1976 : Cainozoic history of southern New Zealand : an accord between geological observations and plate tectonic predictions. Earth and Planetary Science Letters 31. : 85-94.

CHANDLER, M. 1967: "Ceramics in the Modern World." Aldus Books Ltd., London 192p.

COOMBS, D.S. 1960: Lower grade mineral facies in New Zealand. Report XXI International Geological Congress Copenhagen. Part XIII : 339-351.

COPE, R.N. 1965 : The Taranaki Basin : a subsurface review. Shell B.P. and Todd Oil Services Ltd., geological report No. 61 (unpublished).

COPE, R.N.; REED, J.J. 1967: The Cretaceous paleogeography of the Taranaki-Cook Strait area. Proceedings of the Australian Institute of Mining and Metallurgy 222 : 63-72.

CURTIS, C.D.; PETROWSKI, C.; OERTEL, G. 1972 : Stable carbon isotope ratios within carbonate concretions : a clue to place and time of formation. Nature 235: 98-100.

DEER, W.A.; HOWIE, R.A.; ZUSSMAN, J. 1967: Framework Silicates. Volume 4 of "Rock Forming Minerals." Longmans, Green and Co. Ltd., London. 5th Edition. 528p.

DEFFEYES, K.S. 1959: Zeolites in sedimentary rocks. Journal of Sedimentary Petrology 29: 602-609.

DEVEREUX, I. 1967: Oxygen isotope paleotemperature measurements on New Zealand Tertiary fossils. New Zealand Journal of Science 10 : 988-1011.

DICKINSON, W.R. 1971: Detrital modes of New Zealand greywackes. Sedimentary Geology 5 : 37-56.

DUNOYER de SEGONZAC, G. 1970: The transformation of clay minerals during diagenesis and low-grade metamorphism : a review. Sedimentology 15 : 281-346.

EDZWALD, J.K.; O'MELIA, C.R. 1975 : Clay distribution in Recent estuarine sediments. Clays and Clay Minerals 23 : 39-44.

ELPHICK, J.O. 1959: New Zealand oil shale. 4th Triennial Mineral Conference University of Otago Paper 178.

ESWARAN, H. 1972 : Morphology of allophane, imogolite and halloysite. Clay Minerals 9 : 281-286.

FAIRBRIDGE, R.W. 1967 : Phases of diagenesis and authigenesis. Pp. 19-89 in Larsen, G. and Chilingar, G. (Eds): "Diagenesis of Sediments". Elsevier, Amsterdam.

FIELDES, M. 1968: Clay Mineralogy. Pp. 22-39 in Luke, J. (Ed.): Soils of New Zealand. Part 2. New Zealand Soil Bureau Bulletin 26 (2).

FIELDES, M; WILLIAMSON, K.I. 1955 : Clay Mineralogy of New Zealand soils. Part I. Electron Microscopy. New Zealand Journal of Science and Technology B37 : 314-45.

FIELDES, M; McDOWALL, I.C.; CLARIDGE, G.G.C.; WILLIAMS, G.J. 1965: Clays. Pp. 357-370 in Williams G.J. (Ed.) : "Economic Geology of New Zealand." 8th Commonwealth Mining and Metallurgy Congress 4.

FIELDES, M; CLARIDGE, G.; GRAHAM, J. 1968: Soils of New Zealand, Part 3. New Zealand Soil Bureau Bulletin 26 (3). Luke, J. (Ed.).

FINLOW-BATES, T. 1970 : Petrology and Structure of the Metagreywacke Facies Rocks, East of Otorohanga. (Unpublished Ph.D. thesis lodged in the Library, University of Auckland, New Zealand.)

FLEMING, C.A. 1953: The geology of the Wanganui Subdivision. New Zealand Geological Survey Bulletin 52.

FLEMING, C.A. 1962: New Zealand biogeography - a paleontologist's approach. Tuatara 10 : 53-108.

FLEMING, C.A. 1969 : The Mesozoic of New Zealand: chapters in the history of the circum-Pacific mobile belt. Quarterly Journal of the Geological Society of London 125 : 125-170.

FLEMING, C.A. 1975: The geological history of New Zealand and its biota. Pp. 1-86 *in* Kuschel, G. (Ed.) : "Biogeography and Ecology in New Zealand." Dr W. Junk B.V. Publications. The Hague.

FOLK, R.L. 1968: "Petrology of Sedimentary Rocks." Hamphill's, Austin, Texas. 170p.

GIBBS, H.S. 1968: Volcanic soils in New Zealand. *New Zealand Department of Scientific and Industrial Research Information Series 65*.

GIBBS, R.J. 1965: Error due to segregation in quantitative clay mineral X-ray diffraction mounting techniques. *American Mineralogist 50*: 741-751.

GIBBS, R.J. 1968: Clay mineral mounting techniques for X-ray diffraction analysis. *Journal of Sedimentary Petrology 38*: 242-243.

GIBBS, R.J. 1977: Clay mineral segregation in the marine environment. *Journal of Sedimentary Petrology 47*: 237-243.

GLASS, H.D.; POTTER, P.E.; SIEVER, R. 1956: Clay mineralogy of some basal Pennsylvanian sandstones, clays and shales. *Bulletin of the American Association of Petroleum Geologists 40* : 750-754.

GLENN, A.H. and Associates 1973: Meteorological - oceanographic conditions affecting design and operation of offshore tanker terminal facilities: New Plymouth Power Project. New Zealand Electricity Department Report. Unpublished report lodged with Ministry of Works and Development, Wellington. 84pp.

GLENNIE, K.W. 1957: The Lower Miocene formations of Taranaki. *Shell B.P. and Todd Oil Services Ltd. geological report* No. 10 (unpublished).

GLENNIE, K.W. 1959: The graded beds of the Mahoenui Formation (King Country, North Island). *New Zealand Journal of Geology and Geophysics 2*: 613-621.

GOSTIN, V.A.; MORIARTY, K.C. 1975: Preliminary investigation of Tertiary clay mineral distributions around Tasmania, D.S.D.P., Leg 29, *in* Kennett, J.P.; Houtz, R.E. *et al*., Initial Reports of the Deep Sea Drilling Project, Volume 29 : Washington (U.S. Government Printing Office).

GRANT-TAYLOR, T.L. 1964: Volcanic history of western Taranaki. *New Zealand Journal of Geology and Geophysics 7* : 78-86.

GREGG, D.R. 1964: Sheet 18 - Hurunui (1st Ed.) "Geological Map of New Zealand 1:250,000". *Department of Scientific and Industrial Research, Wellington, New Zealand*.

GRIFFIN, G.M. 1962: Regional clay-mineral facies - Products of weathering intensity and current distribution in the north-eastern Gulf of Mexico. *Bulletin of the Geological Society of America 73* : 737-768.

GRIFFIN, J.J.; GOLDBERG, E.D. 1963: Clay mineral distributions in the Pacific Ocean. Pp. 728-741 *in* Hill, M.N. (Ed.): "The Sea." Volume III, New York, Interscience Publication. 963p.

GRIFFITHS, J.R. 1973: Plate tectonics and the evolution of New Zealand. *In* Fraser (Ed.) : "Oceanography of the South Pacific." U.N.E.S.C.O. Publication.

GRIFFITHS, J.R.; VARNE, R. 1972: Evolution of the Tasman Sea, Macquarie Ridge and Alpine Fault. *Nature, Physical Science 235*: 83-86.

GRIM, R.E. 1951 : The depositional environment of red and green shales. *Journal of Sedimentary Petrology* 9 : 226-232.

GRIM, R.E. 1958: "Clay mineralogy." MacGraw - Hill, New York, 384p.

GRIM, R.E.; JOHNS, W.D. 1954: Clay mineral investigations of sediments in the northern gulf of Mexico. Pp. 81-103 *in* Swineford, A. (Ed.) : *Clays and Clay Minerals*, Proceedings of the 2nd National Conference, Pergamon Press, New York.

GRINDLEY, G.W. 1961: Sheet 13 - Golden Bay (1st Ed.) "Geological Map of New Zealand 1:250,000." *Department of Scientific and Industrial Research, Wellington, New Zealand*.

GÜVEN, N. 1974: Lath-shaped units in fine-grained micas and smectites. *Clays and Clay Minerals 22* : 385-390.

HADDOCK, D.L. 1970: Geology of the Tangarakau Gorge - Heao - Tatu Area, North Taranaki. (Unpublished M.Sc. thesis lodged in the Library, University of Auckland, New Zealand.)

HAPPY, A.J. 1971: Tertiary Geology of the Awakino Area. (Unpublished M.Sc. thesis lodged in the Library, University of Auckland, New Zealand.)

HARDER, H. 1972: The role of magnesium in the formation of smectite minerals. *Chemical Geology 10* : 31-40.

HATHERTON, T. 1968: A source for volcanic material in the Mohakatino beds. *New Zealand Journal of Geology and Geophysics 11* : 1207-1210.

HAY, R.F. 1967: Sheet 7 - Taranaki (1st Ed.) "Geological Map of New Zealand 1:250,000." Department of Scientific and Industrial Research, Wellington, New Zealand.

HEALY, J. 1971: The tempo of rhyolitic volcanism in New Zealand. The Acta of the 1st International Scientific Congress on the Volcano of Thera, Athens, Greece. Pp. 64-72.

HEATH, G.R. 1969: Mineralogy of Cenozoic deep-sea sediments from the Equatorial Pacific Ocean. Bulletin of the Geological Society of America 80 : 1997-2018.

HEATH, R.A. 1969: Drift card observations of currents in the central New Zealand region. New Zealand Journal of Marine and Freshwater Research 3 : 3-12.

HEATH, R.A. 1974: Circulation and hydrology of the West Coast of North Island of New Zealand between Cape Terawhiti and Kawhia Harbour. New Zealand Oceanographic Institute Summary No. 6.

HENDERSON, J.; ONGLEY, M. 1923: The geology of the Mokau Subdivision. New Zealand Geological Survey Bulletin 24.

HORNIBROOK, N. de B. 1971: New Zealand Tertiary climate. New Zealand Geological Survey Report 47.

HOVESTADT, H. 1902: Jena glass and its scientific and industrial applications. Macmillan and Company, London, 419p.

HUGHES, R.E.; BOHOR, B.F. 1971 : Scanning electron microscopy of clays and clay minerals. Clays and Clay Minerals 19 : 49-55.

HUME, T.M. 1977: Auckland Thermal No. 1 Investigations. Report on water temperature investigations, southern Manukau Harbour. Unpublished report lodged with Ministry of Works and Development, Auckland. 25pp.

HUME, T.M.; SHERWOOD, A.M.; NELSON, C.S. 1975: Alluvial Sedimentology of the Upper Pleistocene Hinuera Formation, Hamilton Basin, New Zealand. Journal of the Royal Society of New Zealand 5 : 421-462.

IIJIMA, A.; UTADA, M. 1966: Zeolites in sedimentary rocks with reference to the depositional environments and zonal distribution. Sedimentology 7 : 327-357.

JACKSON, M.L.; TYLER, S.A.; WILLIS, A.L.; BOURBEAU, G.A.; PENNINGTON, R.P. 1948: Weathering sequence of clay size minerals in soils and sediments. I. Fundamental generalisations. Journal of Physical Colloid Chemistry 52 : 1237-1260.

JACOBS, M.B. 1970: Clay mineral investigations of Cretaceous and Quaternary deep-sea sediments of the North American Basin. Journal of Sedimentary Petrology 40: 864-868.

JACOBS, M.B.; HAYES, J.D. 1972: Paleo-climatic events indicated by mineralogical changes in deep-sea sediments. Journal of Sedimentary Petrology 42: 889-898.

JACOBS, M.B. 1974: Clay mineral changes in Antartic deep-sea sediments and Cenozoic climatic events. Journal of Sedimentary Petrology 44 : 1079-1086.

JEANS, C.V. 1968: The origin of the montmorillonite of the European Chalk with special reference to the lower chalk of England. Clay Minerals 7 : 311-329.

JOHNS, W.D.; GRIM, R.E.; BRADLEY, W.F. 1954: Quantitative estimations of clay minerals by diffraction methods. Journal of Sedimentary Petrology 24 : 242-251.

JOHNS, W.D.; GRIM, R.E. 1958 : Clay mineral composition of recent sediments from the Mississippi River delta. Journal of Sedimentary Petrology 28: 186-199.

JOHNS, W.D.; SHIMOYAMA, A. 1972: Clay minerals and petroleum forming reactions during burial and diagenesis. Bulletin of the American Association of Petroleum Geologists 56: 2160-2167.

KATZ, H.R. 1968: Potential oil formations in New Zealand and their stratigraphic position as related to basin evolution. New Zealand Journal of Geology and Geophysics 11 : 1077-1133.

KEAR, D. 1960: Sheet 4 - Hamilton (1st Ed.) "Geological Map of New Zealand 1:250,000." Department of Scientific and Industrial Research, Wellington, New Zealand.

KEAR, D.; SCHOFIELD, J.C. 1959: TeKuiti Group, New Zealand Journal of Geology and Geophysics 4 : 148-164.

KEAR, D. 1971: Basement rock facies - northern North Island. New Zealand Journal of Geology and Geophysics 14 : 275-283

KELLER, W.D. 1952: Analcime in the Popo Agie member of the Chugwater formation. Journal of Sedimentary Petrology 22: 70-82.

KELLER, W.D. 1962: "The Principles of Chemical Weathering." Columbia, Mo., Lucas, 111p.

KELLER, W.D. 1968a : Clay minerals as cumulative records of their environments. *Bulletin of the American Association of Petroleum Geologists 52* : 535.

KELLER, W.D. 1968b : Flint clay and flint clay facies. *Clays and Clay Minerals 16* : 113-128.

KELLER, W.D. 1970: Environmental aspects of clay minerals. *Journal of Sedimentary Petrology 40* : 788-813.

KER, D.S. 1973: Stratigraphy and engineering geology of the lower Wanganui Valley, Koriniti to Parakino. *New Zealand Journal of Geology and Geophysics 16*: 189-208.

KING, P.R. 1976: Geological controls on the extent and thickness of the Waikato Coal Measures. *in* Proceedings of the Geological Society of New Zealand Annual Conference, Hamilton, New Zealand. (Abstract).

KIRKMAN, J.H. 1975: Clay mineralogy of some tephra beds of Rotorua area, North Island, New Zealand. *Clay Minerals 10* : 437-449.

KIRKMAN, J.H. 1976: Clay mineralogy of thirteen paleosols developed in Holocene and Late Pleistocene tephras of central North Island, New Zealand. *New Zealand Journal of Geology and Geophysics 19*: 179-87.

KOTELINIKOV, D. 1965: Sur les caractères morphologiques des illites dans les roches sédimentaires. *Mineral Sb. 19* : 26-35.

KRUMBEIN, W.C.; SLOSS, L.L. 1963: "Stratigraphy and Sedimentation." 2nd Edition. W.H. Freeman, San Francisco. 660p.

KUBLER, B. 1966: La cristallinite d'illite et les zones tout à fait supérieur due metomorphisme. Pp. 105-122 *in* : "Colloque sur les Etages Tectoniques à la Baconnière." Neuchatel, Paris.

KUNZE, G. W.; JEFFRIES, C.D. 1953 : X-ray characteristics of clay minerals as related to potassium fixation. *Soil Science 17* : 242-244.

LENSEN, G.J.; FLEMING, C.A.; KINGMA, J.T. 1959: Sheet 10 - Wanganui (1st Ed.) "Geological Map of New Zealand 1:250,000." *Department of Scientific and Industrial Research, Wellington, New Zealand*.

LEWIS, K.B.; EADE, J.V. 1974: Sedimentation in the vicinity of the Maui Gas Field. New Zealand Oceanographic Institute Summary No. 6.

LISITZIN, A.P. 1972: Recent marine clays. Society of Economic Paleotologists and Mineralogists Special Publication 17. 218p.

McCAVE, I.N. 1971 : Wave effectiveness at the sea bed and its relationship to bed-forms and deposition of mud. Journal of Sedimentary Petrology 41: 89-96.

McDOUGALL, J. C. 1961: Ironsand deposits offshore from the West Coast, North Island, New Zealand. New Zealand Journal of Geology and Geophysics 4: 283-300.

McDOUGALL, J.C. 1972: Carbonate variations in slope sediments, Kaipara, New Zealand. New Zealand Journal of Geology and Geophysics 15 : 558-571.

McDOUGALL, J.C.; BRODIE, J.W. 1967: Sediments of the Western shelf, North Island, New Zealand. Memoir of the New Zealand Oceanographic Institute 40. New Zealand Department of Scientific and Industrial Research Bulletin 179. 55p.

McQUILLAN, H. 1977: Hydrocarbon potential of the North Wanganui Basin, New Zealand. The Australian Petroleum Exploration Association Journal 17 : 94-104.

MacEWAN, D.M.C. 1972: Montmorillonite minerals. Pp. 143-207 in Brown, G. (Ed.) : "X-ray Identification and Crystal Structures of Clay Minerals." Mineralogical Society, London. 544p.

MALE, A.G.R. (in press) : Mapping suspended sediment from orbital imagery. Proceedings of the 9th Geographical Society Conference, Dunedin.

MARTIN, K.R. 1967 : "The Mesozoic sequence at south-west Kawhia New Zealand." (Unpublished Ph.D. thesis lodged in the Library, University of Auckland, New Zealand).

MILLOT, G. 1970: "Geology of Clays." Mason et Cie, Paris. 429p.

MILLOT, G.; LUCAS, J.; PAQUET, H. 1966: Evolution chimique par dégradation et agradation des minéraux argileux dans l'hydrosphere. Geol. Rundschau 55 : 1-20.

MILNE, I.H.; EARLY, J.W. 1958: Effect of source and environment on clay minerals. Bulletin of the American Association of Petroleum Geologists 42: 328-338.

MIZUTANI, S. 1970: Silica minerals in the early stages of diagenesis. Sedimentology 15: 419-436.

MOBERLEY, Jr. R.; KIMURA, H.S.; McKOY, F.W. 1968: Authigenic marine phyllosilicates near Hawai. Bulletin of the Geological Society of America 79: 1449-1460.

MOLNAR, P.; ATWATER, R.; MAMMERICKX, J.; SMITH, S.M. 1975: Magnetic anomalies, bathymetry and the tectonic evolution of the South Pacific since the Late Cretaceous. Geophysical Journal of the Royal Astronomical Society 40: 383-420.

MORGENSTEIN, M. 1967: Authigenic cementation of scoriaceous deep-sea sediments west of the Society Ridge, South Pacific. Sedimentology 9: 105-118.

MULLER, G. 1967: Diagenesis in argillaceous sediments. Pp.127-177 in Larsen, G.; Chilingar, G. (Eds.): Diagenesis in Sediments - Developments in Sedimentology 8. Elsevier, Amsterdam.

NEALL, V.E. 1972: Tephrachronology and teprastratigraphy of western Taranaki (N108-109), New Zealand. New Zealand Journal of Geology and Geophysics 15: 507-557.

NEALL, V.E. 1974: "The Volcanic History of Taranaki." Published by Egmont National Park Board.

NELSON, C.S. 1973: Stratigraphy and Sedimentology of the Te Kuiti Group in Waitomo County, South Auckland. (Unpublished Ph.D. thesis lodged in the Library, University of Auckland, New Zealand.)

NELSON, C.S.; COCHRANE, R.H.A. 1970: A rapid x-ray method for the quantitative determination of selected minerals in fine grained and altered rocks. Tane 16 : 152-162.

NELSON, C.S.; HUME, T.M. 1977 : Relative intensity of tectonic events revealed by the tertiary sedimentary record in the North Wanganui Basin and adjacent areas, New Zealand. New Zealand Journal of Geology and Geophysics 20 : 369-392.

NORIN, E. 1953: Occurence of authigenic illitic mica in sediments of the Central Tyrrhenian Sea. Bulletin of the Geological Institute of Uppsala 34 : 279-280.

O'BRIEN, N.R. 1970: The fabric of shale - an electron microscope study. Sedimentology 15 : 229-246.

OERTEL, G.; CURTIS, C.D.; PHAKEY, P.P. 1973: A transmission electron microscope and X-ray diffraction study of muscovite and chlorite. Mineral Magazine 39: 176-88.

PAIN, C.F. 1975 : Some tephra deposits in the southwest Waikato area, North Island, New Zealand. New Zealand Journal of Geology and Geophysics 18 : 541-550.

PARHAM, W.E. 1966: Lateral variations of clay mineral assemblages in modern and ancient sediments. International Clay Conference, 2nd, Jerusalem 1 : 135-146.

PETTIJOHN, J.F. 1975 : "Sedimentary Rocks." (3rd Ed.) Harper and Rowe, Publishers New York. 628p.

PICARD, M.D. 1971 : Classification of fine-grained sedimentary rocks. Journal of Sedimentary Petrology 41: 179-195.

PIERCE, J. W.; SIEGEL, F.R. 1969: Quantification in clay mineral studies of sediments and sedimentary rocks. Journal of Sedimentary Petrology 39 : 187-193.

PORRENGA, D.H. 1966: Clay minerals in Recent sediments of the Niger Delta. Pp. 221-234 in Bailey, S. (Ed.): Clays and Clay Minerals 26, Proceedings of the 14th National Conference, Pergamon Press, New York.

POWERS, M.C. 1957: Adjustment of land derived clays to the marine environment. Journal of Sedimentary Petrology 27 : 355-372.

POWERS, M.C. 1959: Adjustment of land derived clays to chemical change and the concept of equivalence level. Pp. 309-326 in Swineford, A. (Ed.) : Clays and Clay Minerals, Proceedings of the 6th National Conference, Pergamon Press, New York.

PULLAR, W.A.; BIRREL, K.S.; HEINE, J.C. 1973: Named tephras and tephra formations occurring in the central North Island, with notes on derived soils and buried paleosols. New Zealand Journal of Geology and Geophysics 16 : 497-518.

RATEEV, M.A.; GORBUNOVA, Z.N.; LISITZYN, A.P.; NOSOV, G.L.: The distribution of clay minerals in the oceans. Sedimentology 13: 21-43.

REILLY, W.I. 1965: "Gravity Map of New Zealand 1:4,000,000 Bouguer Anomalies." New Zealand Department of Scientific and Industrial Research, Wellington, New Zealand.

REYNOLDS, W.R. 1970: Mineralogy and stratigraphy of Lower Tertiary clays and claystones of Alabama. Journal of Sedimentary Petrology 40 : 823-838.

RICH, C.I.; OBERSHAIN, S.S. 1955 : Chemical and clay mineral properties of a red-yellow podzolic soil derived from muscovite schist. Soil Science 19 : 334-339.

ROSS, G. J. 1969 : Acid dissolution of chlorites: release of magnesium, iron and aliminium and mode of acid attack. Clays and Clay Minerals 17 : 347-354.

ROSS, C.S.; SHANNON, E.V. 1926: Minerals of bentonite and related clays and their physical properties. Journal of the American Ceramics Society 9 : 77-86.

ROSS, C.S.; HENDRICKS, S.B. 1945: Minerals of the Montmorillonite Group. United States Geological Survey Professional Paper 205B.

ROY, R.; SAND, L.B. 1956: A note on some properties of synthetic montmorillonites. American Mineralogist 41 : 505-509.

SARKISYAN, S.G. 1971 : Application of the scanning electron microscope in the investigation of oil and gas resevoir rocks. Journal of Sedimentary Petrology 41 : 289-292.

SARKISYAN, S.G. 1972: Origin of authigenic clay minerals and the significance in petroleum geology. Sedimentary Geology 7 : 1-22.

SAVIN, S.M. 1977 : The history of the earth's surface temperature during the past 100 million years. Annual Review of Earth and Planetary Sciences 5 : 319-355.

SAYLES, F.L.; MANGELSDORF, P.C. 1977: The equilibration of clay minerals with sea water: exchange reactions. Geochimica et Cosmochimica Acta 41: 951-960.

SHERWOOD, A.M. 1973: Surficial Sediments of Raglan Harbour. (Unpublished M.Sc. thesis lodged in the Library, University of Waikato, New Zealand.)

SHOVER, E.F. 1964: Clay-mineral environment relationships in Ciso (U. Penn.) clays and shales, North Central Texas. Pp. 431-443 in Bradley, W. (Ed.): Clays and Clay Minerals, Proceedings of the 12th National Conference, Pergamon Press, New York.

SIEGEL, F. R.; PIERCE, J.W. 1973: Changes in clay mineral assemblages by sampler type. Journal of Sedimentary Petrology 43 : 287-290.

SMOOT, T.W. 1960: Clay mineralogy of pre-Pennsylvanian sandstones and shales of the Illinois Basin, Part 1. - Relation of permeability to clay mineral suites. *Illinois State Geological Survey 286*. 20 pp.

STAINTON, P.W.; GIBSON, G.W. 1964: The geology of central Taranaki. *Shell BP and Todd Oil Services Ltd., geoglical report*. No. 47 (Unpublished).

STOKKE, Per. R.; CARSON, B. 1973: Variation in clay mineral X-ray diffraction results with the quantity of sample mounted. *Journal of Sedimentary Petrology 43* : 957-964.

SYMES, R.E.; WELLS, N. 1973: Mineral content of topsoils on coastal terraces from Mount Egmont to Palmerston North, New Zealand. *New Zealand Journal of Geology and Geophysics 16* : 651-656.

TE PUNGA, M.T. 1964: Relict red-weathered regolith at Wellington. *New Zealand Journal of Geology and Geophysics 7* : 314-339.

TERZAGHI, K. 1966: Influence of geological factors on the engineering properties of sediments. *Economic Geology, 50th Annual Volume* : 557-618.

THOREZ, J. 1975: Phyllosilicates and Clay Minerals. A Laboratory Handbook for their X-ray Diffraction Analysis. Edition G. Lelotte B 4820 Dison, Belgique. 582p.

TOWE, K.M. 1974 : Quantitative clay petrology : the trees but not the forest. *Clays and Clay Minerals 22* : 375-378.

VAN DER LINDEN, W.J.M. 1969: Off-shore sediments, northwest Nelson, South Island, New Zealand. *New Zealand Journal of Geology and Geophysics 15* : 572-589.

VAN DER LINGEN, G.J. 1969: The turbidite problem. *New Zealand Journal of Geology and Geophysics 12*: 7-50.

VAN DER MAREL, H.W. 1954: Potassium fixation in Dutch soils: mineralogical analysis. *Soil Science 78*: 267-287.

VAN DER MERWE, C.R.; WEBER, H.W. 1963: The clay minerals of South African soils developed from granite under different climatic conditions. *South African Journal of Agricultural Science 6* : 411-454.

VELLA, P.; TE PUNGA, M.T.; LIENERT, B.R.; CHRISTOFFEL, D.A. 1976 : Paleomagnetism of two red paleosols and Mangaroa Ash, Upper Hutt, New Zealand. New Zealand Journal of Geology and Geophysics 19 : 849-854.

VUCETICH, C.G.; PULLAR, W.A. 1969: Stratigraphy and chronology of late Pleistocene volcanic ash beds in central North Island, New Zealand. New Zealand Journal of Geology and Geophysics 12. 784-837.

WALKDEN, G.M. 1972: The mineralogy and origin of interbedded clay wayboards in the Lower Carboniferous of the Derbyshire Dome. Geological Journal 8, Part 1 : 143-160.

WATT, D.S. 1965: Natural gas and oil in western New Zealand. Publications of the 8th Commonwealth Mining and Metallurgy Congress 5 : 77-88.

WEAVER, C.E. 1956: The distribution and identification of mixed-layer clays in sedimentary rocks. American Mineralogist 41 : 292-221.

WEAVER, C.E. 1958a: Geologic interpretation of argillaceous sediments. Part I. Origin and significance of clay minerals in sedimentary rocks. Bulletin of the American Association of Petroleum Geologists 42 : 254-271.

WEAVER, C.E. 1958b: Geologic interpretation of argillaceous sediments. Part II. Clay petrology of Mississippian —Lower Pennsylvanian sediments of Central United States. Bulletin of the American Association of Petroleum Geologists 42 : 272-309.

WEAVER, C.E. 1958c: The effects and geological significance of potassium "fixation" by expandable clay minerals derived from muscovite, biotite, chlorite and volcanic material. American Mineralogist 43 : 839-861.

WEAVER, C.E. 1959: The clay petrology of sediments. Pp. 154-187 in Swineford, A. (Ed.) : Clays and Clay Minerals, Proceedings of the 6th National Conference, Pergamon Press, New York.

WEAVER, C.E. 1960: Possible uses of clay minerals in the search for oil. Bulletin of the American Association of Petroleum Geologists 44 : 1505-1518.

WEAVER, C.E.; POLLARD, L.D. 1973: The Chemistry of Clay Minerals. Developments in Sedimentology 15. Elsevier Publ. Co. 213p.

WELLER, J. M. (Ed.) 1960: Glossary of Geology - and related sciences with supplement. (2nd Ed.). American Geological Institute, National Academy of Sciences, National Research Council, Washington, D.C. 545p.

WELLMAN, H. 1973: New Zealand fault zones and sea-floor spreading. in Coleman, P.J. (Ed.) : "The Western Pacific." 335p.

WERMUND, E.G.; MOILA, R.J. 1966: Opal, zeolites and clays in an Eocene neritic sand bar. Journal of Sedimentary Petrology 36: 248-253.

WHITEHOUSE, U.G.; JEFFREY, L.M.; DEBBRECHT, J.D. 1960: Differential settling tendancies of clay minerals in saline waters. Pp. 1-79 in Swineford, A. (Ed.): Clays and Clay Minerals, Proceedings of the 7th National Conference, Pergamon Press, New York.

WILSON, M.D.; PITTMAN, E.D. 1977: Authigenic clays in sandstones: recognition and influence on reservoir properties and paleoenvironmental analysis. Journal of Sedimentary Petrology 47: 3-31.

APPENDIX I

1. Sample : Refers to University of Waikato sample numbers (W.T. series) in cases of onshore, river and harbour samples.

   : Refers to New Zealand Oceanographic Institute sample numbers in cases of western shelf samples.

2. Local. : Sample locality of onshore, river and harbour samples (Fig. 2.1).

3. Strat. : Generalised stratigraphic position of samples.

Rc	Recent
Hw	Hawera
Wc	Castlecliffian
Wn	Nukumaruan
Tg	Tangahoe Mudstone
Mt	Matemateaonga Sandstone
Ur	Urenui Siltstone
Mt.M	Mount Messenger Sandstone
Tw	Tawariki Mudstone
Fy	Ferry Sandstone
Om	Omoao Formation
Pu	Purupuru Volcanic Sandstone
Mg	Mangarara Sandstone
U.Mo	Upper Mokau Sandstone
My	Maryville Coal Measures
L.Mo	Lower Mokau Sandstone
Tmr	Taumarunui Formation
Tmt	Taumatamaire Formation
Ot	Otorohanga Limestone
Wt	Waitomo Sandstone
Or	Orahiri Limestone
Mp	Mangapehi Sandstone
T.A	Te Akatea Siltstone
Ao	Aotea Sandstone
Wh	Whaingaroa Siltstone
G.M	Glen Massey Sandstone
Mk	Mangakotuku Siltstone
Wk	Waikato Coal Measures
Js	Jurassic
Ts	Triassic

4. Lith. : Generalised description of sample lithology.

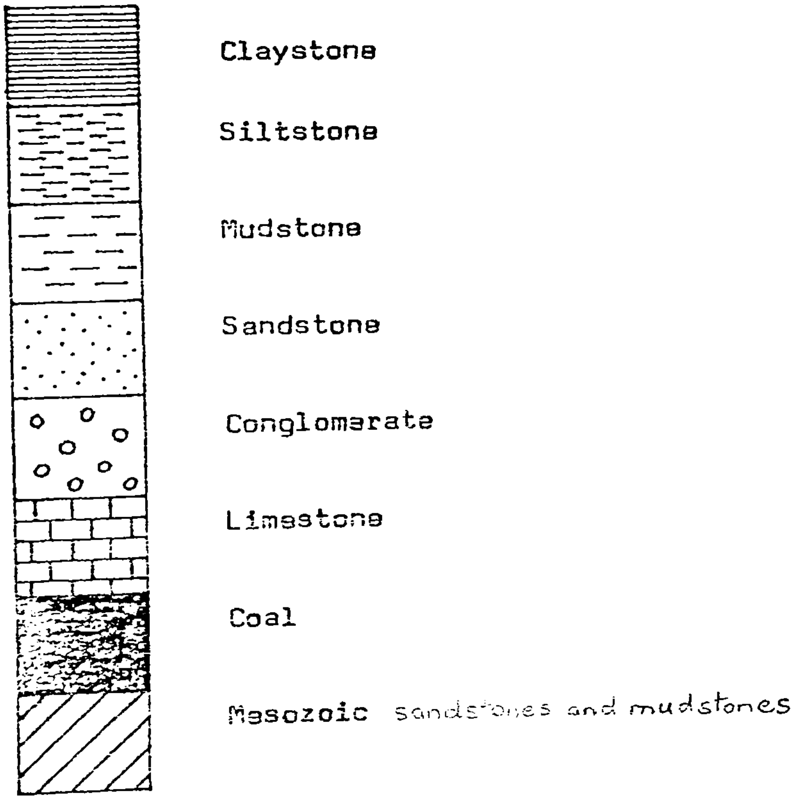
s	soil
md	mud
sd	sand
lst	limestone
mst	mudstone
zst	siltstone
sst	sandstone
cgl	conglomerate

TABLE I.1. Key to abbreviations used in stratigraphic sections (Appendix I) and compositional data sheets (Appendices II to IX).

4. Lith. : Generalised description of sample lithology (continued).

I.mst I.sst	Interbedded mudstone and sandstone units
con	concretion
sty	stylolite
c.	calcareous
carb	carbonaceous

5. Key for symbols used in the stratigraphic sections.



APPENDIX I

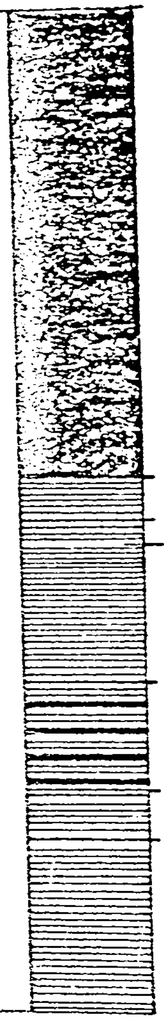
STRATIGRAPHIC SECTIONS

Locality : 1 (N56/694767)

Area : Huntly, McVie Road Mine.

Reference Figure : 1.2

Type Section Area : Waikato Coal Measures



		Thickness in metres	Sample number
Waikato Coal Measures	Black, bituminous coal. Bands with plant roots towards base.	4.0	
	Grey-brown, carbonaceous claystone.	0.3	
	Light grey, carbonaceous claystone with frittered surface.	1.3	10210
	Black bituminous coal in 5-8 cm seams intercolated with light grey bands of carbonaceous claystone.	1.0	
	Grey-brown, carbonaceous claystone.	0.3	
	Light grey, claystone, frittered.	1.4	

TABLE I.2. Stratigraphic and lithologic detail of sediments exposed at localities and sections sampled in the study area. Sample descriptions only are presented for those sites where stratigraphic sections were not erected.

Locality : 2 (N56/586688)

Area: Glen Afton

Sample Description: Mangakotuku Siltstone

10214 : Dark grey, non-calcareous siltstone, weathering ochre brown.

Locality : 3 (N55/546667)

Area: Rotowaro

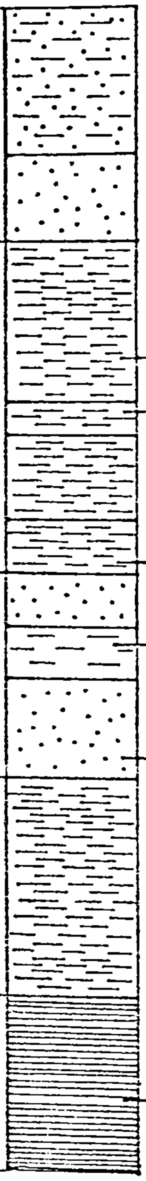
Sample Description: Te Akatea Siltstone

10237 : Brittle, light chocolate-brown, fine-grained very calcareous siltstone.

Locality : 4 (N56/568664)

Area : Te Akatea

Type Section : Te Akatea Siltstone

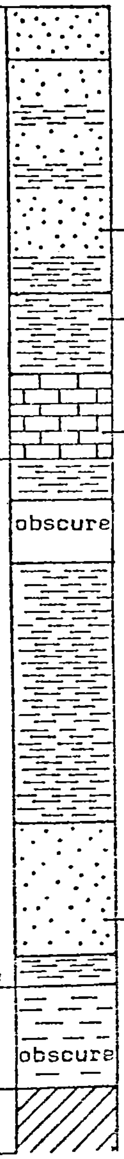


			Thickness in metres	Sample number
Mercer Sandstone		Mottled grey, carbonaceous, silty fine sandstone.	10.0	
		Friable, loose, grey medium sandstone, weathers deep brown.	3.3	
Te Akatea Siltstone		Brittle, light chocolate-brown, fine, very calcareous siltstone with cherty bands.	13.3	10234
		Light grey siltstone.	0.7	10235
		Brittle, light chocolate-brown, fine grained, very calcareous siltstone.	4.2	
		Glauconitic brown siltstone.	4.0	10236
Whaingaroa Siltstone	Aotea Sandstone Facies	Calcareous sandstone.	3.2	
		Flaky, light grey mudstone with concretionary layers.	3.0	10220
		Light grey-brown medium sandstone.	2.0	10223
		Medium blue grey calcareous siltstone.	13.0	
Waikato Coal Measures		Brownish-cream carbonaceous claystone.	2.3	10211

Locality : 5 (N56/597647 - 603643)

Area : Dunphail Bluffs

Type Section : Glen Massey Formation



Formation	Member	Description	Thickness in metres	Sample number
Glen Massey Formation	Glen Massey Sandstone Member	Massive, yellow, calcareous, fine sandstone with extremely rare greywacke pebbles.	7	
		Interbedded, silty fine sandstones and sandy siltstones.	34	10216
	Dunphail Siltstone Member	Blue-grey siltstone, frittered, glauconitic at base.	11	10217
	Elgood Limestone Member	Crystalline, slightly glauconitic limestone with shelly greensand units above and below.	9	10218
Mangakotuku Siltstone		Grey, bedded, calcareous siltstone.	5	
		obscure	7	
		Massive, dark grey siltstone, weathering ochre brown, frittered.	34	
		Medium blue sandstone, fine towards base.	20	10215
		Blue-grey, noncalcareous siltstone with rare fossils.	3	
Waikato Coal Measures		Dark purple brown, carbonaceous mudstone, sandy near top, plant leaves near base.	14	
Mesozoic Basement				

Locality : 6 (see Table I.4.)

Locality : 7 (N64/492463)

Area : Raglan Harbour, shoreline cliffs.

Reference Figure : 1.4

Type Section Area : Whaingaroa Siltstone

Sample Description : Whaingaroa Siltstone

10222 : Frittered, medium blue-gray, calcareous siltstone.

Locality : 8 (N65/583442)

Area : Whatawhata

Sample Descriptions : Soils developed on Mesozoic sandstones and mudstones,

10364 : Yellow clay loam.

10365 : Red clay loam.

Locality : 9 (N64/496428)

Area : Raglan Harbour, Maunurima Creek.

Type Section : Waitetuna Limestone Member of the Aotea Sandstone.

Sample Descriptions : Waitetuna Limestone

10231 : Coarsely flagged, slightly glauconitic, pebbly limestone.

10232 : Finely flagged, sandy, slightly glauconitic, crystalline limestone.

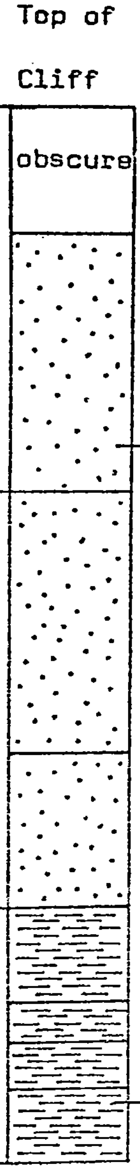
Locality : 10 (see Table I.4.)

Locality : 11 (N73/379199)

Area : Aotea Harbour, shoreline cliffs

Reference Figure : 1.3

Type Section : Aotea Sandstone Formation

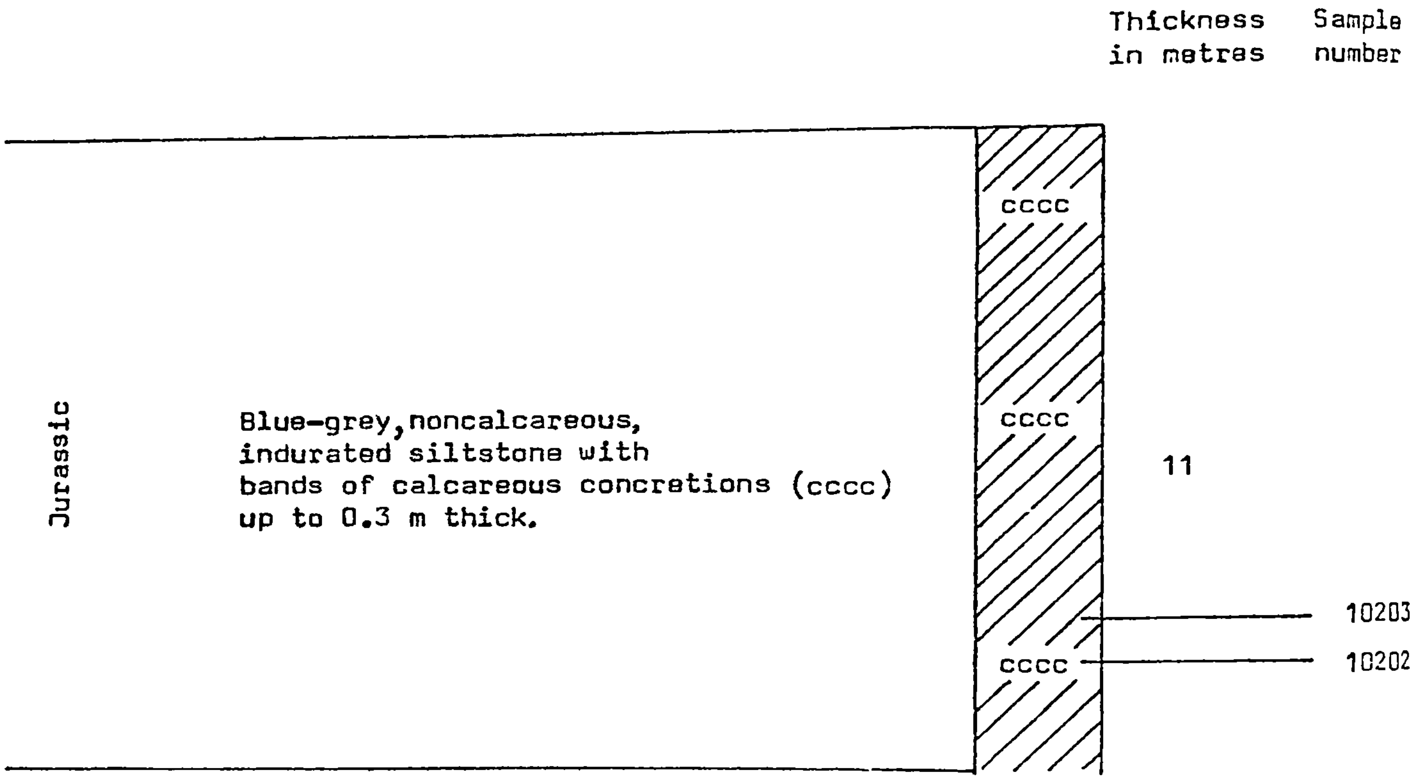


Unit	Description	Thickness in metres	Sample number
	obscure	4.0	
Aotea Sandstone	Steeply bluffed, massive, grey, slightly glauconitic, calcareous, fine sand. Weathers to a yellow-pale brown. Intensively bored at base.	8.0	10227
Gradational contact	Massive, grey, calcareous, fine sand.	8.0	
	Massive, grey, calcareous, fine-very fine sand similar to that above.	5.0	
Whaingaroa siltstone	Frittered, grey, calcareous siltstone.	3.0	
	Blocky, grey, calcareous siltstone.	1.2	
	Frittered, grey, calcareous siltstone.	1.5	10221
	Blocky, grey, calcareous siltstone.	2.0	

Beach

Locality : 12 (N73/405151)

Area : Kawhia, Provincial State Highway 31



Blue-grey, noncalcareous, indurated siltstone with bands of calcareous concretions (cccc) up to 0.3 m thick.

Locality : 13 (N74/846118)

Area : Te Kawa, State Highway 3.

Sample Description : Jurassic mudstone of the Kawhia series

10204 : Slightly weathered, grey-brown, limonite stained, indurated siltstone.

Locality : 14 (N74/657023)

Area : Te Raumauku Hill, roadside outcrop

Sample Description : Aotea Sandstone

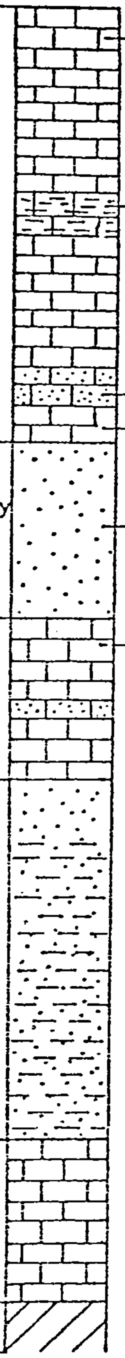
10228 : Moderately weathered, slightly glauconitic, slightly frittered, muddy sandstone.

Locality : 15 (N74/642943 - 640945)

Area : Old Otorohanga Limestone Company Quarry, Waitomo Valley Road.

Reference Figure : 1.7

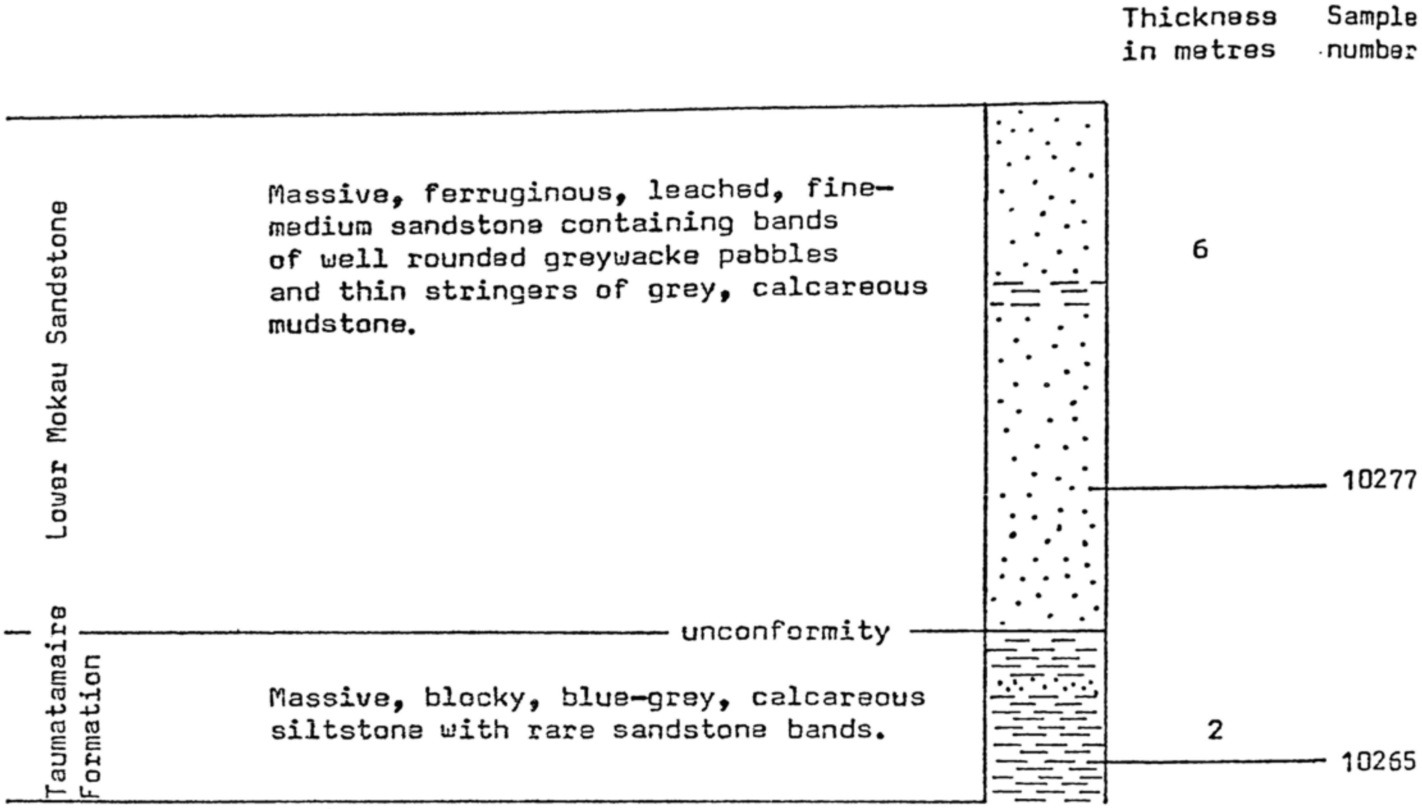
Type Section : Waitomo Sandstone, Orahiri Limestone, Otorohanga Limestone.



		Thickness in metres	Sample number
Otorohanga Limestone	Crystalline limestones that are well-incipiently flagged (10250, 10248) containing cross bedded limestone units, muddy glauconitic limestones (10244), and massive sandy limestone beds (10249).	45	10250 10244 10249 10248
	gradational contact		
Waitomo Sandstone	Massive, calcareous, fine sandstone, commonly leached and weathered. Glauconite and sand content increases upwards.	18	10243
	gradational contact		
Orahiri Limestone	Flaggy crystalline limestone. Irregularly flagged limestone, glauconitic and sandy in places, containing bands of oyster shells up to 60 cm thick.	17	10242
Aotea Sandstone	Top marked by a pebbly fossiliferous, glauconitic, bored, sandy band. Weathered, moderately glauconitic, fossiliferous, muddy sandstone. Passes down into a blue-grey, sandy mudstone.	53	
Waitetuna Limestone	Massive, incipiently seamed, slightly glauconitic sandy limestone, pebbly at top.	17	
	contact not visible		
	Mesozoic basement rocks.		

Locality : 16 (N74/692900)

Area : Hangatiki, National State Highway 3.



Massive, ferruginous, leached, fine-medium sandstone containing bands of well rounded greywacke pebbles and thin stringers of grey, calcareous mudstone.

Massive, blocky, blue-grey, calcareous siltstone with rare sandstone bands.

Locality : 17 (N82/535766)

Area : Mairoa, roadside outcrop.

Sample Descriptions : Mangakotuku Siltstone

Massive, purplish-brown, non-calcareous siltstone (10213), containing calcareous concretions (10212).

Locality : 18 (N82/530766)

Area : Mairoa, roadside outcrop.

Sample Description : Waikato Coal Measures

Creamy-yellow, unconsolidated, carbonaceous mudstones (10207) interbedded with thin carbonaceous beds.

Locality : 19 (N83/673780)

Area : Te Kuiti, National State Highway 3.

Sample : Taumatamaire Siltstone.

10264 : Massive,fossiliferous, blue-grey, calcareous, sandy siltstone.

Locality : 20 (N83/627704)

Area : Te Kuiti, National State Highway 4.

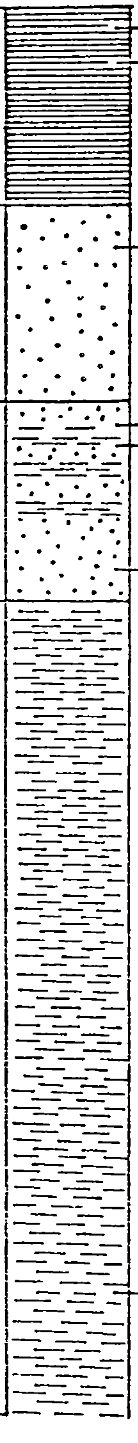
Sample : Taumatamaire Siltstone

10263 : Massive,frittered, fossiliferous, blue-grey, calcareous, siltstone.

Locality : 21 (N83/652687 - 642677)

Area : Te Kuiti, off National State Highway 4.

		Thickness in metres	Sample number
Recent Soils	Red (10367) and white (10366) silty clay soils developed on underlying Tertiary sediments.	17	10367 10366
Upper Mokau Sandstone	Blue-grey, noncalcareous, fine sandstone showing incipient bedding in places.	17	10285
Lower Mokau Sandstone	Interbedded, iron-stained, fossiliferous medium sandstones (10273) and blue-grey non-calcareous siltstones (10272). Chocolate-grey, non-calcareous fine sandstone with carbonaceous laminations.	18	10273 10272 10276
Taumatamairo Formation	Massive, frittered, blue-grey, non-calcareous siltstone. Forms slopes at the base of vertical Mokau Sandstone cliffs.	70	10262



Locality : 22 (N91/540549)

Area : Piopio Serpentine Quarry

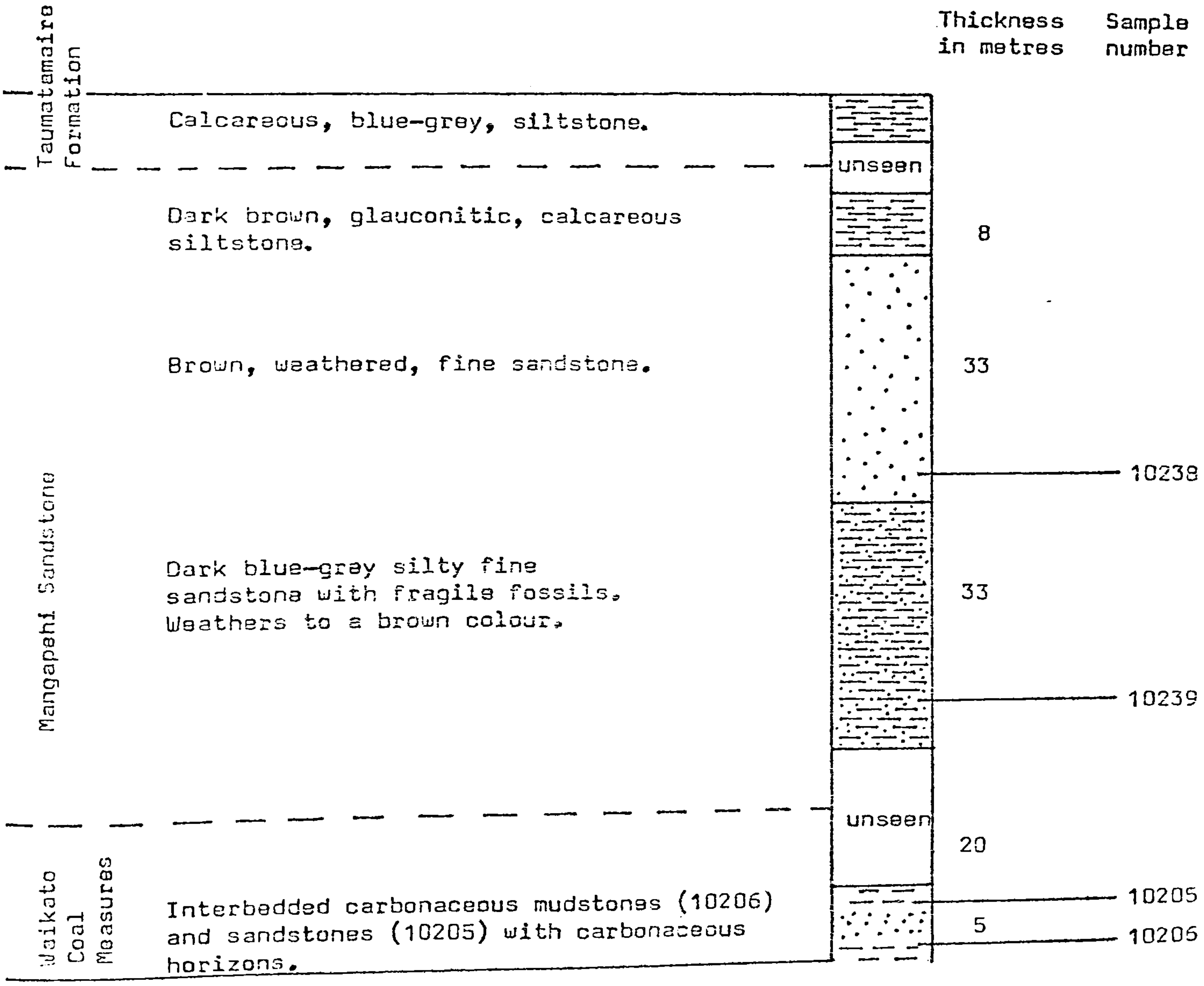
Sample Descriptions : Lower Mokau Sandstone

Massive, brownish, medium sandstone (10278) containing rare calcareous concretions (10274).

Locality : 23 (N92/894587)

Area : Bennedale, Mangepehi Mine.

Type Section : Mangepehi Sandstone



Locality : 24 (N91/532565)

Area : Piopio, Wairere Falls

Sample Descriptions : Otorohanga Limestone

10246 : Flaggy crystalline limestone - Lower Flaggy Limestone.

10245 : Flaggy crystalline limestone - Upper Flaggy Limestone.

Locality : 25 (N92/720521)

Area : Mapiu, National State Highway 4.

Sample Description : Mangarara Sandstone

10292 : Massive, grey, fossiliferous, calcareous, fine sandstone.

Locality : 26 (N91/303472)

Area : Awakino, National State Highway 3.

Reference Figure : 1.9

Sample Descriptions : Taumatamaire Mudstones

Massive, frittered, blue-grey, calcareous, siltstones (10258) with prominent bands of calcareous concretions (10260) up to 60 cm thick.

Locality : 27 (N92/740459)

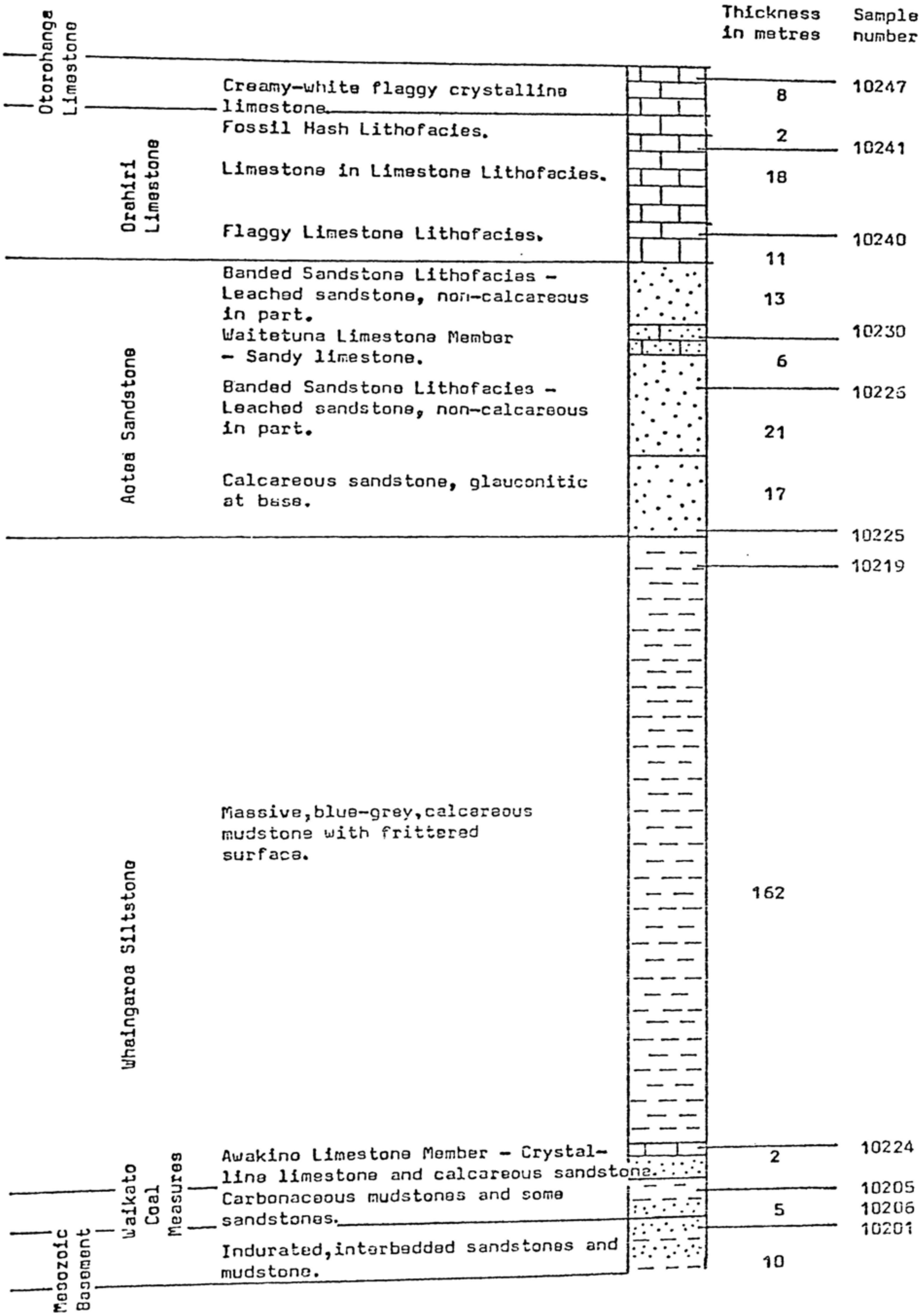
Area : Tangitu, Tangitu Road.

Sample Descriptions : Taumarunui Formation

Blue-grey, calcareous, fine sandstone, with some carbonaceous material, forming beds 5-8 cm thick (10271), alternating with beds of grey, frittered, calcareous, siltstone 15-23 cm thick (10268).

Locality : 28 (N91/274461 – N91/271455)

Area : Awakino Gorge, off National State Highway 3.

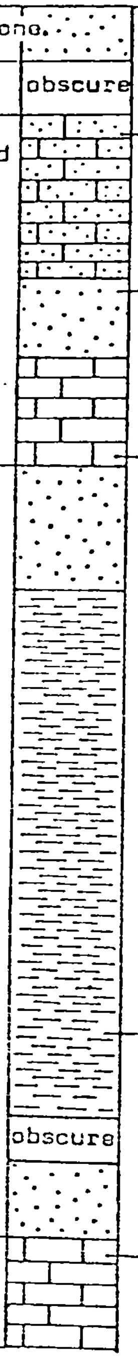


Locality : 29 (see Table I.4.)

Locality : 30 (N91/270427 - N91/274427)

Area : Awakino, Bexley Road.

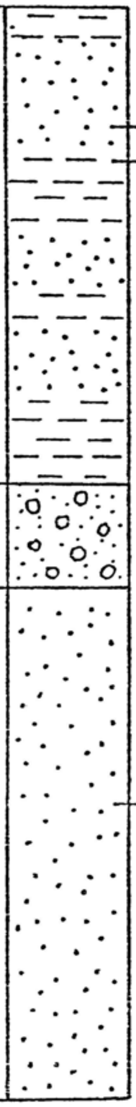
Reference Section : Taumatamaire Formation (Happy, 1971)



Group	Formation	Member	Description	Thickness in metres	Sample number
Mokau Group			Massive, ferruginous, medium sandstone.		
Mahoenui Group	Taumatamaire Formation		obscure	3	
		Upper Awakino Limestone	Hard, slightly sandy, fine-grained limestone. Blocky to flaggy.	10	10255
			Fine, calcareous sandstone (10257) with scattered fossils and concretions (10259).	6	10257 10259
			Blocky, pinkish, fine-grained limestone.	7	10254
			Calcareous fine sandstone with scattered fossils.	8	
			Massive, blue-grey, calcareous siltstone.	45	10261
			obscure		
			Calcareous sandstone.	2	
		Lower Awakino Limestone	Hard, well cemented, pebbly limestone with fragmented fossils.	3	10253

Locality : 31 (N91/165402 - 162402)

Area : Awakino River Mouth.



			Thickness in metres	Sample number
Mohakatino Group	Purupuru Volcanic Sandstone	Interbedded sequence of greenish, non-calcareous sandstone with abundant andesitic volcanic debris (10297) and brownish, non-calcareous micaceous mudstone, with minor volcanic debris (10296).	6	10297 10296
	Mangarara Sandstone	Sandy conglomerate, containing bored sandstone pebbles and cobbles.	0.5	
Mokau Group	Upper Mokau Sandstone	Grey, calcareous, muddy medium sandstone with a few fossils.	5	10289

Locality : 32 (N91/168395)

Area : Awakino, off National State Highway 3.

Sample Descriptions : Omoeo Formation

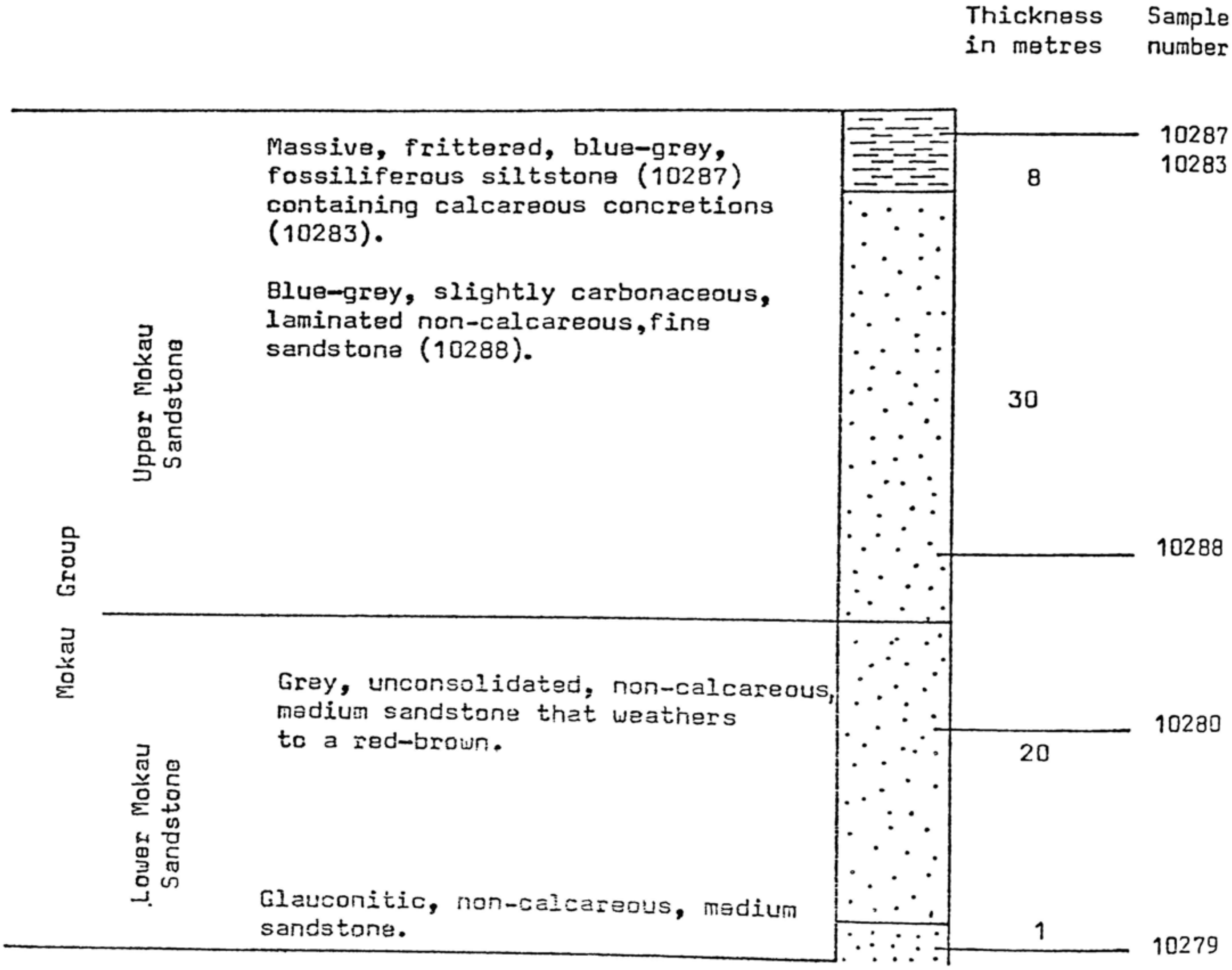
10300 : Blue-grey, calcareous siltstone with minor volcanic debris.

10302 : Grey, calcareous, micaceous, fine sandstone with volcanic debris interbedded with (10300).

10303 : Massive, porous, brown-green, micaceous, medium sandstone.

Locality : 33 (N91/188407 – 187403)

Area : Awakino, off National State Highway 3.



Massive, frittered, blue-grey, fossiliferous siltstone (10287) containing calcareous concretions (10283).

Blue-grey, slightly carbonaceous, laminated non-calcareous, fine sandstone (10288).

Grey, unconsolidated, non-calcareous, medium sandstone that weathers to a red-brown.

Glauconitic, non-calcareous, medium sandstone.

Locality : 34 (N91/249386)

Area : Awakino, Awakau Road

Sample Descriptions : Upper Awakino Limestone

10252 : Creamy-white, flaggy limestone rich in calcareous algae.

10251 : Blue-grey, terrigenous seam material from between limestone flags.

Locality : 35 (see Table 1.4.)

Locality : 36 (N91/160350)

Area : Mokau River Mouth

Sample Descriptions : Ferry Sandstone

Grey, micaceous, calcareous, medium sandstone with some volcanic debris, forming beds 3 cm thick (10301), alternating with beds of blue-grey calcareous siltstone, forming beds 1 m thick (10299) and containing rare concretions (10298).

Locality : 37 (see Table 1.4.)

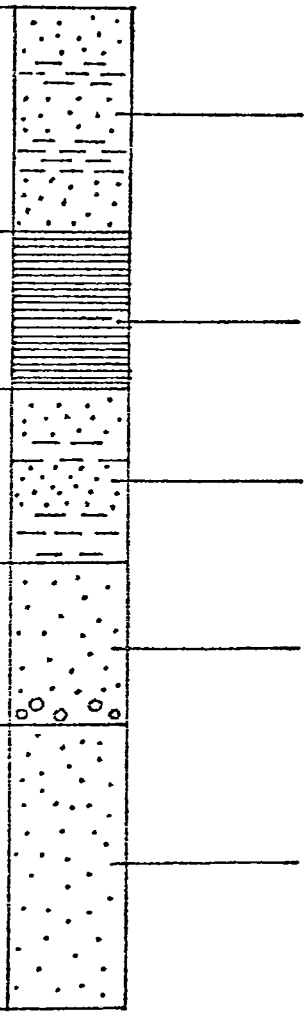
Locality : 38 (N100/207288 - N91/184300)

Area : Mohakatino Road

Reference Figure : 1.17

Type Section : Mohakatino Group

			Sample number
Mohakatino Group	Ferry Sandstone	Greenish, micaceous, tuffaceous sandstone (10311), interbedded with blue-grey, tuffaceous, siltstone (10307).	10311 10307
	Upper Purupuru Volcanic Sandstone	Non-calcareous, claystone, containing volcanic debris.	10294
	Lower Purupuru Volcanic Sandstone	Grey, calcareous, medium sandstone containing abundant volcanic debris (10295) and tuffaceous mudstone.	10295
	Mangarara Sandstone	Grey, calcareous, medium sandstone containing abundant volcanic debris Thin basal conglomerate.	10293
Mokau Group		Blue-grey, non calcareous, carbonaceous fine sandstone.	10286



Locality : 39 (N101/702248)

Area : Okahukura, Opoitiki Road

Sample Description : Ferry Sandstone

10312 : Green, micaceous, tuffaceous sandstone.

Locality : 40 (see Table 1.4.)

Locality : 41 (N101/737205)

Area : Taumarunui, National State Highway 4.

Sample Descriptions : Ferry Sandstone.

Laminated, brownish-green, micaceous, fine sandstone (10310) interbedded with bands of massive, grey-green, micaceous sandy siltstone (10306).

Locality : 42 (N100/150170)

Area : Ahititi, National State Highway 3.

Reference Figure : 1.19

Sample Description : Mt. Messenger Sandstone

10317 : Massive, blue-grey, micaceous, medium sandstone.

Locality : 43 (N100/267157)

Area : Kotare, Provincial State Highway 40.

Sample Description : Ferry Sandstone

Blue-grey, non-calcareous, micaceous medium sandstone (10308) interbedded with blue-grey, micaceous, mudstone (10304) with rare fossils.

Locality : 44 (N100/318168)

Area : Waitaanga, Provincial State Highway 40.

Sample Description : Upper Mokau Sandstone

10284 : Massive blue-grey, calcareous, siltstone.

Locality : 45 (N101/755158)

Area : Taumarunui, National State Highway 4.

Type Section Area : Taumarunui Formation

Sample Descriptions : Taumarunui Formation

Grey, calcareous, sandstone (10270) interbedded with blue-grey, calcareous siltstone (10276).

Locality : 46 (N100/133110)

Area : Uruti, National State Highway 4.

Type Section Area : Mt. Messenger Sandstone

Sample Descriptions :

10318 : Brownish, micaceous, non-calcareous medium sandstone - Mt. Messenger Sandstone

10323 : Blue-grey, micaceous, slightly calcareous siltstone with some carbonaceous material - Urenui Siltstone.

10320 : Calcareous concretion from 10323.

Locality : 47 (N100/447103)

Area : Ohura, Tatu Coal Mine.

Sample Description : Lower Mokau Sandstone

10281 : Creamy-brown, carbonaceous, non-calcareous, sandy mudstone.

Locality : 48 (N100/446093)

Area : Ohura, Waro Road, Tatu Mine.

Sample Descriptions : Upper Mokau Sandstone

10290 : Green, carbonaceous, sandy siltstone.

10291 : Green-brown sandy conglomerate.

Locality : 49 (see Table I.4.)

Locality : 50 (N99/032041)

Area : Uruti, National State Highway 3.

Reference Figure : 1.21

Type Section Area : Urenui Siltstone

Sample Description : Urenui Siltstone

Blue-grey, micaceous, calcareous siltstone (10324), with thin tuffaceous sandstone bands (10327) and concretionary bands (10321).

Locality : 51 (N100/451036)

Area : Tangarakau Gorge, Provincial State Highway 43.

Sample Description : Tawariki Mudstone

10315 : Massive, blue-grey, micaceous, non-calcareous siltstone.

10315 : Calcareous concretion from 10315.

Locality : 52 (see Table I.4.)

Locality : 53 (see Table I.4.)

Locality : 54 (N100/356004)

Area : Tangarakau Gorge, Provincial State Highway 43.

Sample Description : Lower Mokau Sandstone

10275 : Massive, blue-grey, non-calcareous medium sand.

Locality : 55 (see Table I.4.)

Locality : 56 (N110/337914)

Area : Tahora Saddle, Provincial State Highway 43.

Sample Description : Ferry Sandstone

Brown-green, micaceous, non-calcareous, medium sand (10309) interbedded with blue-grey, micaceous, non-calcareous siltstone (10305).

Locality : 57 (N110/286855)

Area : Whangamomona, Provincial State Highway 43.

Sample Description : Tawariki Mudstone

10316 : Massive, blue-grey, slightly calcareous mudstone with sparse fossils.

10314 : Calcareous concretion from 10316.

Locality : 58 (N111/855874)

Area : National Park, National State Highway 4.

Reference Figure : 1.10

Sample Description : Taumarunui Formation

Flysch-type sequence of grey, slightly calcareous fine sandstones (10269), interbedded with blue-grey calcareous siltstones (10266).

Locality : 59 (N110/219793 – 214793)

Area : Pohokura, Provincial State Highway 43.

Sample Descriptions :

10319 : Massive, brown-grey, micaceous, medium sand –
– Mt. Messenger Sandstone.

10325 : Massive, blue-grey, micaceous, siltstone
– Urenui siltstone.

10322 : Concretion from 10325.

Locality : 60 (N110/176775)

Area : Pohokura Saddle, Provincial State Highway 43.

Sample Description : Urenui Siltstone

10326 : Massive, blue-grey, micaceous, non-calcareous siltstone.

Locality : 61 (see Table 1.4.)

Locality : 62 (N110/121708)

Area : Te Wera, Provincial State Highway 43.

Sample Description : Matemateaonga Sandstone

10331 : Massive, blue-grey, micaceous, fine sandstone with rare fossils.

Locality : 63 (N119/922610)

Area : Stratford, Standish Road

Sample Description : Matemateaonga Sandstone

10333 : Massive, friable, blue-grey, micaceous, non-calcareous sandstone.

Locality : 64 (N119/057637)

Area : Strathmore Saddle, Provincial State Highway 43.

Reference Figure : 1.22

Sample Descriptions : Matemateaonga Sandstone

10332 : Massive or incipiently bedded, blue-grey, fossiliferous fine sandstone.

10328 : Calcareous concretion from 10332.

Locality : 65 (N120/121507)

Area : Matemateaonga, Rawhitiroa Road.

Type Section Area : Matemateaonga Sandstone

Sample Descriptions : Matemateaonga Sandstone

10334 : Massive, blue-grey, non-calcareous fine sandstone.

10329 : Calcareous concretion from 10334.

Locality : 66 (see Table I.4)

Locality : 67 (N121/712448)

Area : Raetihi, Pipiriki Road.

Sample Descriptions : Matemateaonga Sandstone

10335 : Massive green-grey, micaceous, non-calcareous, medium sandstone.

10330 : Calcareous concretion from 10335.

Locality : 68 (N136/945341)

Area : Hawera, Tangahoe Valley Road.

Type Section Area : Tangahoe Mudstone

Sample Description : Tangahoe Mudstone

10341 : Massive, blue-grey, fossiliferous, mudstone.

Locality : 69 (N131/796284)

Area : Oreore, National State Highway 4.

Sample Descriptions : Tangahoe Mudstone

Brown-grey, micaceous, non-calcareous, medium sandstone units averaging 7 m thick (10349) interbedded with blue-grey, micaceous, non-calcareous, mudstone units averaging 1.2 m. thick (10348).

Locality : 70 (N129/914240)

Area : Tangahoe River, at National State Highway 3.

Sample Descriptions : Tangahoe Mudstone

10340 : Massive, blue-grey, micaceous, non-calcareous mudstone.

10336 : Calcareous concretion from 10340.

Locality : 71 (N132/244239)

Area : Taihape, National State Highway 1.

Reference Figure : 1.23

Sample Description : Tangahoe Mudstone

10347 : Massive, frittered, micaceous, slightly calcareous, sandy mudstone.

Locality : 72 (N131/838206)

Area : Kakatahi, National State Highway 4.

Reference Figure : 1.24

Sample Descriptions : Tangahoe Mudstone.

10344 : Massive, blue-grey, slightly calcareous, sandy mudstone.

10337 : Calcareous concretion from 10344.

Locality : 73 (N132/254206)

Area : Taihape, National State Highway 1.

Sample Description : Tangahoe Mudstone

10339 : Calcareous concretion from Tangahoe Mudstone.

Locality : 74 (N136/047074)

Area : Patea River Mouth at National State Highway 3.

Sample Description : Tangahoe Mudstone

10342 : Massive, blue-grey, non-calcareous, micaceous mudstone.

Locality : 75 (N139/303082)

Area : Utiki, National State Highway 1.

Sample Descriptions : Tangahoe Mudstone

10346 : Massive, blue-grey, micaceous, calcareous sandy mudstone.

10338 : Calcareous concretion from 10346.

Locality : 76 (N137/285016)

Area : Waitotara River at State Highway 3.

Sample Descriptions : Tangahoe Mudstone

10343 : Massive - poorly bedded, grey, micaceous, non-calcareous mudstone.

Locality : 77 (N138/726034)

Area : Parahauhau Junction, National State Highway 4.

Sample Description : Tangahoe Mudstone

10345 : Massive, blue-grey, micaceous, calcareous sandy mudstone.

Locality : 78 (N138/772040)

Area : Otomoa, National State Highway 4.

Sample Description : B horizon of soil developed on Tangahoe mudstone.

10369 : Brown, micaceous, silty loam.

Locality : 79 (N138/659968)

Area : Upokongaro, National State Highway 4.

Sample Descriptions : Interbedded carbonaceous mudstones, siltstones, sands and pebbly conglomerates with numerous fossils - age Nukumaruan.

10352 : Weathered, micaceous, non-calcareous, fine-medium sandstone.

10351 : Grey-brown, micaceous, non-calcareous siltstone.

10350 : Chocolate-brown, carbonaceous, micaceous, non-calcareous, muddy siltstone.

Locality : 80 (N139/225976)

Area : Ohingaiti, National State Highway 1.

Sample Descriptions : Nukumaruan Sediments

10353 : Poorly bedded, fossiliferous, dark-grey, micaceous, silty, fine sandstones.

Locality : 81 (N138/617908)

Area : Wanganui, National State Highway 4.

Sample Descriptions : Shell beds, conglomerates, current bedded sands and muds of Castlecliffian age.

10360 : Rhyolitic, pumiceous, fossiliferous, coarse sands.

10357 : Brown sandy mud.

Locality : 82 (see Table 1.4.)

Locality : 83 (N139/063849)

Area : Hunterville, National State Highway 1.

Sample Descriptions : Castlecliffian Sediments

10356 : Pale-grey, poorly bedded, unfossiliferous, micaceous, fine sandstone.

Locality : 84 (N139/019808)

Area : Hunterville, National State Highway 1.

Sample Descriptions : Castlecliffian Sediments

Interbedded brown, micaceous, non-calcareous fine sands (10355) and grey, non-calcareous silts (10354).

Locality : 85 (N143/690783)

Area : Whanagehu, National State Highway 3.

Sample Descriptions : Lower Hawera Sediments - interbedded marine andesitic and "greywacke" gravels, pumiceous sands and clays.

10358 : brown mud.

10361 : brown, pumiceous, fine sand.

Locality : 86 (N143/738718)

Area : Turakina, National State Highway 3.

Sample Descriptions : Hawera Sediments - Interbedded andesitic sand, sandy silts and muds.

10363 : Brown-grey, slightly micaceous andesitic, fine - medium sand.

10359 : Chocolate-red, carbonaceous, mud.

10362 : Brown, micaceous, silty sand.

Locality : 87 (see Table 1.4.)

Locality : 88 (N91/502343)

Area : Waitewhena, Open Cast Mine.

Sample Description : Maryville Coal Measures

10282 : Creamy-brown, carbonaceous, silty mudstone.

APPENDIX I

Sample Number	Date of Collection	Location	Position	
			Latitude	Longitude
N377	12.12.74	Hamilton Shelf	37^{o} 48.5'	174^{o} 45.8'
N378	"	" "	37^{o} 48.9'	174^{o} 39.4'
N379	"	" "	37^{o} 48.9'	174^{o} 13.2'
N382	13.12.74	Central Taranaki Shelf	39^{o} 15'	173^{o} 42.4'
N383	"	"	39^{o} 15.7'	173^{o} 42.6'
N384	"	"	39^{o} 15.9'	173^{o} 39.4'

TABLE I.3. Western shelf seston sample data.

APPENDIX I.

SAMPLE DATA : RIVER AND HARBOUR SEDIMENTS

Sample number	Locality number	Locality	Grid reference	Sampling gear	Stratigraphy	Sediment description
10396	87	Rangitikei River	N148/743448	Sc	Rc	Grey-brown, slightly micaceous, sandy mud.
10395	82	Wangenui River	N138/578848	Sc	Rc	Grey-green slightly micaceous muddy sand.
10394	76	Waitotara River	N137/285016	G	Rc	Grey, micaceous, sandy mud.
10393	74	Patea River	N136/047074	Sc	Rc	Grey, micaceous, sandy mud.
10392	74	Patea River	N136/047074	Sc	Rc	Grey micaceous mud.
10391	55	Waiwakaiho River	N109/681928	Sc	Rc	Grey-brown, coarse sand, rich in ferromagnesian minerals.
10390	61	Waroa River	N118/408693	Sc	Rc	" " " " " " " "
10389	66	Waiaua River	N118/462428	Sc	Rc	Grey, sandy gravel, rich in andesitic rock fragments.
10388	70	Tangahoe River	N129/914240	Sc	Rc	Green-grey, micaceous, muddy sand.
10387	52	Waitara River	N109/806992	Sc	Rc	Green-brown, micaceous, silty mud.
10386	53	Urenui River	N109/923996	Sc	Rc	Grey, slightly micaceous, sandy mud.
10385	49	Mimi River	N99/005035	Sc	Rc	Greenish-brown, slightly micaceous, mud.
10384	40	Tongaporutu River	N100/142214	Sc	Rc	Greenish-brown, slightly micaceous, sandy mud.
10383	37	Mohakatino River	N91/159320	Sc	Rc	Greenish-brown, slightly micaceous, silty mud.
10382	35	Mokau River	N91/170359	Sc	Rc	Greenish-brown, sandy mud, rich in ferromagnesian minerals.
10381	29	Awakino River	N91/170417	Sc	Rc	Greenish-brown, silty mud.
10380	6	Raglan Harbour	N64/398473	G	Rc	Greenish-brown, sandy mud, containing ferromagnesian minerals.
10379	6	"	N64/445457	G	Rc	Greenish-brown, mud, containing shell fragments.
10378	6	"	N64/411452	G	Rc	Greenish-brown, sandy mud.
10377	6	"	N55/467510	G	Rc	Brown mud.
10376	12	Kawhia Harbour	N73/405151	Sc	Rc	Green-brown sandy mud.
10375	12	"	N73/342123	Sc	Rc	Greyish muddy sand.
10374	10	Aotea Harbour	N73/373181	Sc	Rc	Green-brown sandy mud.
10373	10	"	N64/347204	Sc	Rc	Grey muddy sand containing ferromagnesian minerals.

TABLE I.4. Sample localities and field descriptions, river and harbour sediments. Sampling gear: Sc = Scoop, G = Grab sampler.

APPENDIX II

Thesis Numbers		Field Numbers		Thesis Numbers		Field Numbers	
Locality	Sample	Locality	Sample	Locality	Sample	Locality	Sample
67	10396	88	RB.1	85	10361	49	49.2
82	10395	89	RB.2	81	10360	50	50.1
76	10394	57	RB.3	86	10359	48	48.2
74	10393	58	RB.4A	85	10358	49	49.1
74	10392	58	RB.4B	81	10357	50	50.2
55	10391	74	RB.5A	83	10356	46	46.1
61	10390	75	RB.6A	84	10355	47	47.1
66	10389	76	RB.7A	84	10354	47	47.2
70	10388	77	RB.8A	80	10353	45	45.1
52	10387	78	RB.9	79	10352	51	51.1
53	10386	79	RB.10	79	10351	51	51.2
49	10385	80	RB.11	79	10350	51	51.3
40	10384	84	RB.12	69	10349	55	55.1
37	10383	85	RB.13	69	10348	55	55.2
35	10382	86	RB.14	71	10347	43	43.1
29	10381	87	RB.15	75	10346	44	44.1
6	10380	90	90.1	77	10345	52	52.1
6	10379	90	90.2	72	10344	54	54.1
6	10378	90	90.3	76	10343	57	57.1
6	10377	90	90.4	74	10342	58	58.1
12	10376	91	91.1	68	10341	59	59.1
12	10375	91	91.3	70	10340	77	77.1
10	10374	92	92.1	73	10339	43	43.2
10	10373	92	92.2	75	10338	44	44.2
55	10372	74	RB.5B	72	10337	54	54.2
61	10371	75	RB.6B	70	10336	77	77.2
66	10370	76	RB.7B	67	10335	56	56.1
73	10369	53	53.1	65	10334	60	60.1
70	10368	77	RB.8B	63	10333	61	61.1
21	10367	35	35.5A	64	10332	62	62.1
21	10366	35	35.5B	62	10331	63	63.1
3	10365	30	30.1	67	10330	56	56.2
8	10364	30	30.2	65	10329	60	60.2
86	10363	48	48.1	64	10328	62	62.2
86	10362	48	48.3	50	10327	81	81.2

TABLE II.1. Correlation of thesis sample localities and sample numbers with those used in the course of field work. All thesis sample numbers have prefix W.T.

TABLE II.1. (continued).

Thesis Numbers		Field Numbers		Thesis Numbers		Field Numbers	
Locality	Sample	Locality	Sample	Locality	Sample	Locality	Sample
60	10326	64	64.1	48	10291	71	71.2
59	10325	65	65.2	48	10290	71	71.1
50	10324	81	81.1	31	10289	18	18.2
46	10323	82	82.2	33	10288	16	16.4
59	10322	65	65.3	33	10287	16	16.9
50	10321	81	81.3	38	10286	20	20.1
46	10320	82	82.3	21	10285	35	35.4
59	10319	65	65.1	44	10284	72	72.1
46	10318	82	82.1	33	10283	16	16.10
42	10317	83	83.1	88	10282	88	10282
57	10316	66	66.1	47	10281	70	70.1
51	10315	69	69.1	33	10280	16	16.14
57	10314	66	66.2	33	10279	16	16.15
51	10313	69	69.2	22	10278	25	25.1
39	10312	39	39.1	16	10277	32	32.1
38	10311	20	20.4A	21	10276	35	35.2
41	10310	41	41.1A	54	10275	68	68.1
56	10309	67	67.1	22	10274	25	25.2
43	10308	73	73.1	21	10273	35	35.3A
38	10307	20	20.4B	21	10272	35	35.3B
41	10306	41	41.1B	27	10271	38	38.1
56	10305	67	67.2	45	10270	40	40.1A
43	10304	73	73.2	58	10269	42	42.1A
32	10303	17	17.2	27	10268	38	38.2
32	10302	17	17.3B	45	10267	40	40.1B
36	10301	19	19.1	58	10266	42	42.1B
32	10300	17	17.3A	16	10265	32	32.2
36	10299	19	19.2	19	10264	33	33.1
36	10298	19	19.3	20	10263	34	34.1
31	10297	18	18.1A	21	10262	35	35.1
31	10296	18	18.1B	30	10261	21	21.5
38	10295	20	20.3A	26	10260	23	23.2
38	10294	20	20.3B	30	10259	21	21.5a
38	10293	20	20.2	26	10258	23	23.1
25	10292	37	37.1	30	10257	21	21.2

TABLE II.1. (continued).

Thesis Numbers		Field Numbers		Thesis Numbers		Field Numbers	
Locality	Sample	Locality	Sample	Locality	Sample	Locality	Sample
30	10256	21	21.4	14	10228	2	2.3
30	10255	21	21.1	11	10227	10	10.6
30	10254	21	21.3	28	10226	22	22.5
30	10253	21	21.6	28	10225	22	22.4
34	10252	26	26.2	28	10224	22	22.2
34	10251	26	26.1	4	10223	11	11.1
15	10250	1	1.4	7	10222	4	4.1
15	10249	1	1.12	11	10221	10	10.1
15	10248	1	1.13	4	10220	11	11.2
28	10247	22	22.12	28	10219	22	22.3
24	10246	24	24.1	5	10218	6	6.1
24	10245	24	24.3	5	10217	6	6.2
15	10244	1	1.9	5	10216	6	6.3
15	10243	1	1.15	5	10215	6	6.13
15	10242	1	1.17	2	10214	7	7.1
28	10241	22	22.7	17	10213	29	29.1
28	10240	22	22.9	17	10212	29	29.2
23	10239	36	36.2	4	10211	11	11.9
23	10238	36	36.3	1	10210	12	12.4
3	10237	8	8.1	28	10209	22	22.1A
4	10236	11	11.3	28	10208	22	22.1B
4	10235	11	11.5	18	10207	28	28.2
4	10234	11	11.6	23	10206	36	36.1B
15	10233	1	1.25	23	10205	36	36.1A
9	10232	5	5.1	13	10204	31	31.1
9	10231	5	5.2	12	10203	91	91.2A
28	10230	22	22.6	12	10202	91	91.2B
15	10229	1	1.21	28	10201	22	22.17

APPENDIX III. BULK MINERAL COMPOSITION

ONSHORE SAMPLES

Sample	Local.	Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
10372	55	Rc	s	-	97	-	-	-	3	-	97	-	-	3 All
10371	61	Rc	s	15	47	-	-	-	38	15	47	-	-	38 All
10370	66	Rc	s	-	52	-	-	-	48	-	52	-	-	48 All
10369	78	Rc	s	28	15	-	40	1	16	28	15	-	40	17
10368	70	Rc	s	28	43	5	33	-	-	28	43	5	33	-
10367	21	Rc	s	37	-	-	75	-	-	37	-	-	75	-
10366	21	Rc	s	38	-	-	80	-	-	38	-	-	80	-
10365	8	Rc	s	17	-	-	85	-	-	17	-	-	85	-
10364	8	Rc	s	21	4	2	75	-	-	21	4	2	75	-
10363	86	Hw	sd	15	42	-	-	-	43	15	42	-	-	43
10362	86	Hw	sd	14	39	6	30	1	10	14	39	6	30	11
10361	85	Hw	sd	22	43	6	25	1	3	22	43	6	25	4
10360	81	Hw	sd	14	90	-	9	6	-	15	96	-	10	-
10359	86	Hw	md	21	41	6	20	2	10	21	42	6	20	11
10358	85	Hw	md	15	23	-	49	2	11	15	23	-	50	12
10357	81	Hw	md	22	18	-	37	3	20	23	19	-	38	20
10356	83	Wc	sst	24	21	3	35	1	16	24	21	3	35	17
10355	84	Wc	I.sst	22	23	3	55	1	-	22	23	3	55	-
10354	84	Wc	I.mst	17	18	3	65	2	-	17	18	3	66	-
10353	80	Wn	c.sst	20	18	4	28	11	19	25	23	5	35	12
10352	79	Wn	sst	30	16	-	35	1	18	30	16	-	35	19
10351	79	Wn	mst	32	18	3	30	-	17	32	18	3	30	17
10350	79	Wn	mst	29	22	4	48	1	-	29	22	4	48	-

TABLE III.1. Bulk mineral compositions of samples R.Qtz. = % quartz; R. Plag. = % Plagioclase feldspar; R.Pot. = % potash feldspar; R.Clay = % clay minerals obtained by XRD modal analysis of acid insoluble residues, $CaCO_3$ calculated following acid digestion, Qtz. = % quartz, etc., obtained after recalculating insoluble residue analyses so as to include $CaCO_3$. Other minerals present include allophane (All), analcite (An) and α-cristobalite (Cs).

TABLE IV.1. (continued).

Sample	Local.	Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Othors	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
10349	69	Tg	I.sst	46	36	-	15	1	2	46	36	-	15	3
10348	69	Tg	I.mst	22	27	-	57	1	-	22	27	-	57	-
10347	73	Tg	mst	21	22	5	36	4	12	22	23	5	38	12
10346	75	Tg	mst	40	39	7	33	5	-	43	42	7	35	-
10345	77	Tg	mst	30	21	-	32	3	14	31	22	-	34	13
10344	72	Tg	mst	25	20	-	55	1	-	25	20	-	55	-
10343	76	Tg	mst	22	30	3	35	1	9	22	30	3	35	10
10342	74	Tg	mst	19	21	4	51	3	2	20	22	4	53	1
10341	68	Tg	mst	19	29	3	42	2	4	20	30	3	43	4
10340	70	Tg	mst	20	26	4	30	1	19	20	26	4	30	20
10339	73	Tg	con	20	15	4	15	27	19	28	20	5	20	27
10338	75	Tg	con	26	22	6	21	29	-	37	31	8	30	-
10337	72	Tg	con	11	14	-	33	32	10	16	21	-	48	15
10336	70	Tg	con	18	18	-	28	20	16	22	22	-	35	21
10335	67	Mt	sst	28	38	5	25	1	3	28	38	5	25	4
10334	65	Mt	sst	37	30	7	35	-	-	37	30	7	35	-
10333	63	Mt	sst	28	29	3	47	1	-	28	29	3	47	-
10332	64	Mt	sst	21	28	4	36	5	6	22	30	4	38	6
10331	62	Mt	sst	28	32	-	30	1	9	28	32	-	30	10
10330	67	Mt	con	13	17	5	14	43	8	22	30	9	25	14
10329	65	Mt	con	17	20	4	21	40	-	21	34	7	35	3
10328	64	Mt	con	22	32	-	27	28	-	30	45	-	38	-
10327	50	Ur	sst	2	21	4	10	3	60	2	22	4	10	63 An
10326	60	Ur	mst	21	29	3	48	1	-	21	29	3	48	-
10325	59	Ur	mst	25	23	-	48	1	3	25	23	-	48	4
10324	50	Ur	mst	25	31	-	35	1	8	25	31	-	35	9
10323	46	Ur	mst	20	28	-	41	3	8	21	29	-	43	7
10322	59	Ur	con	14	19	-	21	38	8	22	31	-	34	13
10321	50	Ur	con	19	16	-	21	29	15	27	22	-	30	9
10320	46	Ur	con	14	22	2	20	29	13	20	31	3	28	18

TABLE III.1. (continued).

Sample	Local.	Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others	
10319	59	Mt.M	sst	40	33	7	15	1	4	40	33	7	15	5	
10318	46	Mt.M	sst	35	29	-	30	1	5	35	29	-	30	6	
10317	42	Mt.M	sst	30	34	4	28	1	3	30	34	4	28	4	
10316	57	Tw	mst	23	24	4	46	3	-	24	25	4	48	-	
10315	51	Tw	mst	31	25	4	37	-	3	31	25	4	37	3	
10314	57	Tw	con	12	18	-	30	30	10	17	26	-	43	14	
10313	51	Tw	con	19	17	3	28	30	3	27	24	4	40	5	
10312	39	Fy	sst	25	35	-	44	1	-	25	35	-	45	-	
10311	38	Fy	I.sst	28	43	7	19	-	3	28	43	7	19	3	
10310	41	Fy	I.sst	35	17	5	38	-	5	35	17	5	38	5	
10309	56	Fy	I.sst	30	31	3	35	-	1	30	31	3	35	1	
10308	43	Fy	I.sst	37	24	7	42	-	-	37	24	7	42	-	
10307	38	Fy	I.mst	22	22	-	50	-	6	22	22	-	50	6	
10306	41	Fy	I.mst	30	32	-	30	-	8	30	32	-	30	8	
10305	56	Fy	I.mst	16	24	3	40	1	16	16	24	3	40	17	
10304	43	Fy	I.mst	23	28	-	46	2	1	23	28	-	47	2	
10303	32	Om	sst	32	37	2	27	-	2	32	37	2	27	2	An
10302	32	Om	I.sst	3	53	-	8	18	18	4	64	-	10	22	An
10301	36	Om	I.sst	22	41	-	15	24	-	29	54	-	20	-	
10300	32	Om	I.mst	7	10	-	25	36	2	15	23	-	56	6	
10299	36	Om	I.mst	28	19	-	51	8	-	30	21	-	56	-	
10298	36	Om	con	18	20	3	31	34	-	28	31	4	47	-	
10297	31	Pu	I.sst	2	78	4	-	8	8	2	85	4	-	9	Cs
10296	31	Pu	I.mst	16	19	-	41	2	22	16	19	-	42	23	
10295	38	Pu	sst	11	38	3	26	10	12	12	42	3	29	14	Cs, An
10294	38	Pu	mst	14	11	-	85	-	-	14	11	-	85	-	
10293	38	Mg	c.sst	12	19	-	49	12	8	14	22	-	56	8	
10292	25	Mg	c.sst	28	17	8	20	20	7	35	21	10	25	9	

TABLE III.1. (continued).

Sample	Local.	Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
10291	48	U.Mo	cgl	37	49	6	28	-	-	37	49	6	28	-
10290	48	U.Mo	sst	38	30	3	38	-	-	38	30	3	38	-
10289	31	U.Mo	c.sst	12	9	2	25	45	7	22	17	3	46	12
10288	33	U.Mo	sst	29	18	-	47	-	6	29	18	-	47	6
10287	33	U.Mo	mst	32	29	-	27	3	9	33	30	-	28	9
10286	38	U.Mo	mst	28	28	-	46	-	-	28	28	-	46	-
10285	21	U.Mo	mst	35	33	5	33	1	-	35	33	5	33	-
10284	44	U.Mo	mst	35	27	5	38	1	-	35	27	5	38	-
10283	33	U.Mo	con	16	15	3	18	51	-	33	31	7	37	-
10282	88	My	mst	40	5	-	55	-	-	40	5	-	55	-
10281	47	L.Mo	mst	41	12	5	40	-	2	41	12	5	40	2
10280	33	L.Mo	sst	36	24	4	33	-	3	36	24	4	33	3
10279	33	L.Mo	sst	23	24	6	40	-	7	23	24	6	40	7
10278	22	L.Mo	sst	32	28	3	29	-	8	32	28	3	29	8
10277	16	L.Mo	sst	43	34	5	28	-	-	43	34	5	28	-
10276	21	L.Mo	sst	47	32	-	30	1	-	47	32	-	30	-
10275	54	L.Mo	sst	47	32	12	28	-	-	47	32	12	28	-
10274	22	L.Mo	con	9	7	2	18	59	5	23	18	5	45	9
10273	21	L.Mo	I.sst	31	14	5	35	1	14	31	14	5	35	15
10272	21	L.Mo	I.mst	25	23	-	47	1	4	25	23	-	47	5
10271	27	Tmr	I.sst	28	20	-	35	7	10	30	22	-	38	10
10270	45	Tmr	I.sst	27	22	-	26	24	1	36	29	-	34	1
10269	58	Tmr	I.sst	37	22	5	44	3	-	38	23	5	45	-
10268	27	Tmr	I.mst	15	18	-	64	4	-	16	19	-	67	-
10267	45	Tmr	I.mst	19	18	-	53	3	7	20	18	-	55	-
10266	58	Tmr	I.mst	18	14	3	64	6	-	21	15	3	68	-

TABLE III.1. (continued).

Sample	Local.	Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
10265	16	Tmt	c.mst	11	11	-	45	29	-	15	15	-	64	6
10264	19	Tmt	c.mst	8	7	3	41	42	-	13	12	5	70	-
10263	20	Tmt	c.mst	14	13	3	60	20	-	17	16	4	75	-
10262	21	Tmt	mst	27	12	-	62	1	-	27	12	-	63	-
10261	30	Tmt	c.mst	28	21	-	43	8	-	31	23	-	47	-
10260	26	Tmt	c.mst	20	18	-	40	12	10	23	20	-	45	12
10259	30	Tmt	con	17	11	-	27	43	2	30	20	-	47	3
10258	26	Tmt	con	16	13	-	33	27	11	22	18	-	45	15
10257	30	Tmt	c.sst	19	20	4	39	15	3	22	24	5	46	3
10256	30	Tmt	c.sst	19	11	2	28	41	-	32	18	4	47	-
10255	30	Tmt	lst	1	1	-	2	97	-	17	14	4	51	14
10254	30	Tmt	lst	1	1	-	2	96	-	28	17	5	56	-
10253	30	Tmt	lst	5	4	-	13	76	2	21	18	-	56	5
10252	34	Tmt	lst	4	3	1	10	82	-	20	16	4	56	4
10251	34	Tmt	sty	9	15	4	20	48	4	17	28	7	38	10
10250	15	Ot	lst	1	-	-	2	97	-	11	4	2	55	28
10249	15	Ot	lst	4	2	-	4	90	-	36	15	3	37	9
10248	15	Ot	lst	3	1	-	2	96	-	45	16	-	32	5
10247	28	Ot	lst	6	1	1	2	90	-	57	6	8	16	13
10246	24	Ot	lst	-	-	-	2	98	-	17	9	14	47	13
10245	24	Ot	lst	-	-	-	2	98	-	25	14	6	41	14
10244	15	Ot	c.sst	5	4	-	18	74	-	19	17	-	70	-
10243	15	Wt	c.sst	20	8	3	41	24	-	28	11	4	57	-
10242	15	Or	lst	4	8	-	7	82	-	25	45	-	38	-
10241	28	Or	lst	12	12	8	5	66	-	35	35	24	14	-
10240	28	Or	lst	8	8	7	6	82	-	42	58	38	20	-
10239	23	Mp	c.sst	47	16	5	29	23	-	61	21	7	38	-
10238	23	Mp	sst	60	16	7	35	-	-	60	16	7	35	-

TABLE III.1. (continued).

Sample	Local.	Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others	
10237	3	T.A	mst	5	2	2	40	37	14	10	4	4	65	27	Cs
10236	4	T.A	mst	24	5	-	32	42	-	41	8	-	56	-	
10235	4	T.A	mst	15	6	-	35	47	-	28	11	-	66	-	
10234	4	T.A	mst	20	8	7	26	44	-	35	15	12	47	-	
10233	15	Ao	lst	17	7	4	3	71	-	58	25	12	10	-	
10232	9	Ao	lst	3	1	-	6	91	-	31	10	-	60	-	
10231	9	Ao	lst	7	16	2	17	56	2	16	36	4	38	6	
10230	28	Ao	lst	7	5	5	2	79	2	32	24	23	9	12	
10229	15	Ao	c.sst	18	9	7	32	33	-	27	14	10	47	2	
10228	14	Ao	c.sst	10	7	5	25	53	-	21	15	11	47	6	
10227	11	Ao	c.sst	14	12	5	23	47	-	26	22	10	44	-	
10226	28	Ao	c.sst	28	13	22	-	48	-	53	25	42	-	-	
10225	28	Ao	sst	50	12	29	37	-	-	50	12	29	37	-	
10224	28	Wh	lst	1	2	-	3	95	-	17	22	7	65	-	
10223	4	Wh	c.sst	21	4	-	28	50	-	42	9	-	56	-	
10222	7	Wh	c.mst	16	7	-	33	44	-	28	13	-	58	1	
10221	11	Wh	c.mst	15	11	7	21	50	-	31	22	14	42	-	
10220	4	Wh	c.mst	18	6	-	37	42	-	31	11	-	64	-	
10219	28	Wh	c.mst	12	7	2	36	44	-	21	12	4	65	-	
10218	5	G.M	lst	1	1	-	5	90	3	10	6	3	48	33	
10217	5	G.M	c.mst	14	2	4	25	46	9	26	4	8	46	16	
10216	5	G.M	c.sst	17	-	8	20	41	14	29	2	14	34	21	
10215	5	Mk	mst	24	6	-	75	-	-	24	6	-	75	-	
10214	2	Mk	mst	30	4	5	59	8	-	32	4	5	62	1	
10213	17	Mk	mst	37	3	5	75	-	-	37	3	5	75	-	
10212	17	Mk	con	9	1	1	18	72	-	32	3	3	65	-	

TABLE III.1. (continued).

Sample	Local.	Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
10211	4	Wk	mst	37	-	-	140	-	-	37	-	-	140	-
10210	1	Wk	mst	52	-	-	120	-	-	52	-	-	120	-
10209	28	Wk	mst	22	-	-	150	-	-	22	-	-	150	-
10208	28	Wk	mst	45	9	3	120	-	-	45	9	3	120	-
10207	18	Wk	mst	46	-	4	110	-	-	46	-	4	110	-
10206	23	Wk	mst	75	6	-	57	-	-	75	6	-	57	-
10205	23	Wk	sst	74	-	-	30	-	-	74	-	-	30	-
10204	13	Js	mst	22	25	8	20	-	25	22	25	8	20	25
10203	12	Js	mst	22	13	-	52	1	12	22	13	-	52	13
10202	12	Js	con	3	9	-	25	63	-	8	25	-	66	1
10201	28	Ts	sst	13	37	5	20	-	25	13	37	5	20	25

RIVER AND HARBOUR SAMPLES

Sample	Local.	Strat.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others	
10396	87	Rc	28	27	5	34	-	6	28	27	5	34	6	
10395	82	Rc	35	27	-	28	-	10	35	27	-	28	10	
10394	57	Rc	45	40	8	26	-	-	45	40	8	26	-	
10393	58	Rc	21	42	6	38	-	-	21	42	6	38	-	
10392	58	Rc	30	38	5	28	-	-	30	38	5	28	-	
10391	55	Rc	-	36	-	10	-	54	-	36	-	10	54	All
10390	61	Rc	15	47	-	-	-	38	15	47	-	-	38	All
10389	66	Rc	-	100	-	-	-	-	-	100	-	-	-	
10388	70	Rc	24	41	5	24	-	6	24	41	5	24	6	
10387	52	Rc	26	28	-	46	-	-	26	28	-	46	-	
10386	53	Rc	27	38	3	28	-	4	27	38	3	28	4	
10385	49	Rc	28	20	-	42	-	10	28	20	-	42	10	
10384	40	Rc	24	35	4	28	1	8	24	35	4	28	19	
10383	37	Rc	35	26	4	25	1	9	35	26	4	25	10	
10382	35	Rc	14	19	4	38	1	24	14	19	4	30	25	
10381	29	Rc	28	26	4	24	-	8	28	26	4	24	8	

TABLE III.1. (continued).

Sample	Local.	Strat.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
10380	6	Rc	15	13	2	54	6	10	16	14	2	57	11
10379	6	Rc	15	7	2	42	2	32	15	7	2	43	33
10378	6	Rc	23	17	3	52	2	-	23	17	3	53	-
10377	6	Rc	20	11	-	73	-	-	20	11	-	73	-
10376	12	Rc	21	13	4	37	3	22	22	13	4	28	23
10375	12	Rc	7	22	7	19	3	42	7	23	7	20	43
10374	10	Rc	27	25	7	10	2	29	28	26	7	10	29
10373	10	Rc	8	28	3	23	3	35	8	29	3	24	36

WESTERN SHELF SAMPLES

Sample	Strat.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz	R.Plag.	R.Pot.	R.Clay	Others
A45	Rc	18	19	-	26	12	25	20	22	-	30	27
A47	Rc	12	20	5	25	17	21	15	24	6	30	25
A51	Rc	17	19	-	39	25	-	22	25	-	52	1
B319	Rc	24	24	5	29	23	-	31	31	7	38	-
B321	Rc	16	17	2	42	19	4	20	21	3	52	4
B322	Rc	20	33	5	30	1	1	20	33	5	30	2
B645	Rc	12	18	2	46	22	-	15	23	3	42	1
B647	Rc	33	18	7	27	11	4	37	20	8	30	5
B648	Rc	35	20	6	23	7	9	38	22	6	25	9
B649	Rc	12	17	2	43	10	16	13	19	2	48	18
B653	Rc	2	11	-	9	10	68	2	12	-	10	24
B667	Rc	7	41	7	10	2	33	7	42	7	10	34
B675	Rc	26	23	4	31	9	7	29	25	4	34	8
B676	Rc	27	30	-	25	1	17	27	30	-	25	18
B680	Rc	16	29	8	37	3	7	16	30	8	38	8
B681	Rc	28	27	4	29	2	10	28	27	4	29	12
B682	Rc	22	20	3	33	21	1	28	25	4	42	1
B778	Rc	16	17	4	20	20	23	20	21	5	25	21
B780	Rc	15	13	3	15	23	31	20	17	4	19	40
B781	Rc	17	18	4	12	22	27	22	23	5	15	35
B784	Rc	33	31	6	24	12	-	37	35	7	27	-

TABLE III.1. (continued).

Sample	Strat.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
B788	Rc	21	32	4	8	15	20	25	38	5	10	22
B793	Rc	40	25	5	21	9	-	43	27	5	23	2
B795	Rc	41	34	11	13	12	-	47	39	12	15	-
B796	Rc	22	22	4	16	27	9	30	30	5	22	13
B799	Rc	14	32	5	10	1	38	14	32	5	10	39
B801	Rc	17	26	7	38	1	11	17	26	7	38	12
B802	Rc	14	21	3	27	2	33	14	21	3	28	34
B804	Rc	11	31	-	13	15	30	13	37	-	15	35
B812	Rc	16	23	3	-	2	56	16	23	3	-	58
B813	Rc	13	22	5	37	2	21	13	22	5	38	22
B816	Rc	16	22	4	31	18	9	20	27	5	38	10
B818	Rc	27	22	5	22	23	1	35	29	7	28	1
B822	Rc	35	37	8	25	1	-	35	37	8	25	-
C169	Rc	14	17	4	32	29	4	20	24	5	45	6
C170	Rc	27	35	8	27	2	1	28	36	8	28	-
C171	Rc	16	20	5	29	2	28	16	20	5	30	29
C172	Rc	15	18	-	54	2	11	15	18	-	55	12
C173	Rc	14	24	3	39	2	18	14	24	3	40	19
C174	Rc	22	27	11	39	2	-	22	27	11	40	-
C175	Rc	14	33	-	37	4	12	15	34	-	38	13
C178	Rc	14	22	12	28	2	22	14	22	12	28	24
C262	Rc	20	39	4	9	8	20	22	42	4	10	22
C265	Rc	20	22	6	20	10	22	22	24	7	22	25
C272	Rc	26	23	7	24	16	4	31	27	8	28	6
C275	Rc	30	28	5	19	20	-	37	35	6	24	-
C276	Rc	41	18	6	22	8	5	44	20	7	24	5
C277	Rc	38	25	5	23	8	1	42	28	6	25	-
C281	Rc	21	29	4	14	6	26	22	31	4	15	28
C293	Rc	24	26	5	13	13	19	28	30	6	15	21
C330	Rc	23	19	3	14	7	34	25	21	3	15	36
C332	Rc	27	33	9	15	3	13	28	34	9	15	14
C340	Rc	18	25	2	14	15	26	21	29	2	17	41
C348	Rc	24	37	-	22	4	13	25	38	-	22	15
C355	Rc	29	41	7	25	9	-	32	45	8	28	-

TABLE III.1. (continued).

Sample	Strat.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$	Others	R.Qtz.	R.Plag.	R.Pot.	R.Clay	Others
C363	Rc	35	19	7	19	6	14	37	20	8	20	15
C364	Rc	38	25	5	–	3	29	40	25	5	–	30
C372	Rc	30	27	6	12	19	6	37	33	8	15	7
C376	Rc	15	41	5	17	7	15	15	42	5	18	20
C380	Rc	11	41	11	14	4	9	12	43	12	15	18
C421	Rc	14	25	6	33	11	11	16	28	7	37	12
C429	Rc	14	19	5	39	19	4	17	24	6	48	5
C432	Rc	27	18	3	18	14	20	32	21	4	22	21
C434	Rc	16	22	3	38	21	–	20	27	4	48	1
C435	Rc	16	29	–	42	2	11	16	29	–	43	12
C436	Rc	14	17	2	43	18	6	17	21	3	52	7
C438	Rc	30	19	5	14	28	4	42	27	7	20	4
C450	Rc	19	30	–	14	5	32	20	32	–	15	33

APPENDIX III. BULK MINERAL COMPOSITION

ONSHORE SAMPLES

Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$
Hw	sd	16	53	3	16	2
Hw	md	19	27	2	35	2
Wc	sst	24	21	3	35	1
Wc	I.sst	22	23	3	55	1
Wc	I.mst	17	18	3	65	2
Wn	c.sst	25	17	2	31	6
Wn	mst	31	20	4	39	1
Tg	I.sst	46	36	–	15	1
Tg	I.mst	22	27	–	57	1
Tg	mst	24	26	3	39	2
Mt	sst	28	31	4	35	2
Ur	sst	2	21	4	10	3
Ur	mst	23	28	1	43	2
Mt.M	sst	35	32	4	24	1
Tw	mst	27	24	4	41	2
Fy	sst	25	35	–	44	1
Fy	I.sst	32	29	5	33	–
Fy	I.mst	23	26	1	41	1
Om	sst	32	37	2	27	–
Om	I.sst	12	47	–	11	21
Om	I.mst	17	10	–	38	32
Pu	I.sst	2	78	4	–	8
Pu	I.mst	16	19	–	41	2
Pu	mst	11	38	3	26	10
Pu	mst	14	11	–	85	–
Mg	c.sst	20	19	4	35	16
U.Mo	cgl	37	49	6	28	–
U.Mo	sst	26	19	2	37	15
U.Mo	mst	32	29	2	36	1
L.Mo	mst	41	12	5	40	–
L.Mo	sst	38	29	5	31	–
L.Mo	I.sst	31	14	5	35	1
L.Mo	I.mst	25	23	–	47	1

TABLE III.2. Average bulk mineral composition of onshore samples (formational average) and river and harbour samples (locality average), Qtz. = % quartz; Plag. = % plagioclase feldspar; Pot. = % potash feldspar; Clay = % clay minerals; obtained by XRD modal analysis. $CaCO_3$ = % $CaCO_3$ calculated following acid digestion.

TABLE III.2. (continued).

Strat.	Lith.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$
Tmr	I.sst	31	21	2	35	11
Tmr	I.mst	17	17	1	60	4
Tmt	c.mst	20	15	1	47	17
Tmt	c.sst	19	16	3	33	28
Tmt	lst	3	2	–	7	88
Ot	lst	2	–	–	2	96
Ot	c.sst	5	4	–	18	74
Wt	c.sst	20	8	3	41	24
Or	lst	8	9	5	6	77
Mp	c.sst	53	16	6	32	11
T.A	mst	16	5	2	33	42
Ao	lst	8	7	3	7	74
Ao	c.sst	35	11	20	27	21
Wh	lst	1	2	–	3	95
Wh	c.sst	21	4	–	28	50
Wh	c.mst	15	8	2	32	45
G.M	lst	1	1	–	5	90
G.M	c.mst	14	2	4	25	46
G.M	c.sst	17	–	8	20	41
Mk	mst	30	4	3	70	3
Wk	mst	46	2	1	116	–
Wk	sst	74	–	–	30	–
Js	mst	22	19	4	36	1
T.s	sst	13	37	5	20	–

RIVER AND HARBOUR SAMPLES

Local.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$
87	28	27	5	34	–
82	35	27	–	28	–
57	45	40	8	26	–
58	25	40	6	33	–
55	–	36	–	10	–
61	15	47	–	–	–
66	–	100	–	–	–
70	24	41	5	24	–
52	26	28	–	46	–
53	27	38	3	28	–
49	28	20	–	42	–
50	24	35	4	28	1

TABLE III .2. (continued).

Local.	Qtz.	Plag.	Pot.	Clay	$CaCO_3$
37	35	26	4	25	1
35	14	19	4	38	1
29	28	26	4	24	–
6	18	12	2	55	2
12	14	17	5	28	3
10	17	27	5	16	3

APPENDIX IV. CLAY FRACTION COMPOSITION

ONSHORE SAMPLES

Sample	Local.	Strat.	Lith.	K	I	M	C		M/L	M/L Species						Others
							xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10372	55	Rc	s	–	–	–	–	–	–	–	–	–	–	–	–	P, All
10371	61	Rc	s	–	–	–	–	–	–	–	–	–	–	–	–	P, All
10370	66	Rc	s	–	–	–	–	–	–	–	–	–	–	–	–	P, All
10369	78	Rc	s	8	72	8	–	3	9	–	–	*	9	–	–	Q, P
10368	70	Rc	s	3	36	5	–	9	47	5	12	*	30	–	–	Q, P
10367	21	Rc	s	69	30	1	–	–	–	–	–	–	–	–	–	–
10366	21	Rc	s	32	68	–	–	–	–	–	–	–	–	–	–	Q
10365	8	Rc	s	58	42	–	–	–	–	–	–	–	–	–	–	–
10364	8	Rc	s	24	76	–	–	–	–	–	–	–	–	–	–	Q
10363	86	Hw	sd	–	46	48	–	–	6	6	–	–	–	–	–	Q, P
10362	86	Hw	sd	–	51	36	–	–	13	4	–	*	9	–	–	Q, P
10361	85	Hw	sd	–	39	58	–	–	3	3	–	–	–	–	–	Q, P
10360	81	Hw	sd	4	44	43	–	–	9	–	–	*	9	–	–	
10359	86	Hw	mst	–	32	68	–	–	–	–	–	–	–	–	–	P
10358	85	Hw	mst	3	39	58	–	–	–	–	–	–	–	–	–	P
10357	81	Hw	mst	4	44	43	–	–	9	–	–	*	9	–	–	Q, P, k
10356	83	Wc	sst	6	44	37	–	7	6	–	–	*	6	–	–	Q, P
10355	84	Wc	I.sst	3	47	43	–	–	7	–	–	*	7	–	–	P
10354	84	Wc	I.mst	5	49	44	–	–	2	–	–	*	2	–	–	P

TABLE IV.1. Clay fraction ($<2\mu$) composition of samples as determined by XRD analysis, K = % kaolinite; I = % illite; M = % montmorillonite; C = total chlorite %; xC = % crystalline chlorite; hC = % heat-labile chlorite; M/L = total mixed-layer clay mineral %; I-M = % illite-montmorillonite mixed-layer clay; C-I = % chlorite-illite mixed-layer clay; C-V = % chlorite-vermiculite mixed-layer clay; C-M = % chlorite-montmorillonite mixed-layer clay (*distinction between C-V and C-M not possible; C-I-M = % chlorite-illite-montmorillonite mixed-layer clay; U.M/L = % undifferentiated mixed-layer clay; Q = quartz; P = plagioclase feldspar; K = potash feldspar; 2P = plagioclase feldspars A and B both present in subequal amounts; An = analcite; Am = amphibole; Cs = α-cristobalite and All = allophane. N.B. K + I + M + C + M/L = 100%.

TABLE IV.1. (continued).

Sample	Local.	Strat.	Lith.	K	I	M	C		M/L	M/L Species						Others
							xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10353	80	Wn	c.sst	-	52	23	-	12	13	-	-	*	13	-	-	P, k
10352	79	Wn	sst	2	45	50	-	-	3	-	-	*	3	-	-	Q, P
10351	79	Wn	mst	2	44	32	15	-	7	-	-	-	-	7	-	Q, P, k
10350	79	Wn	mst	2	42	32	-	20	4	-	-	-	-	4	-	Q, P, k
10349	69	Tg	I.sst	1	45	32	-	4	18	-	9	*	9	-	-	Q, P, k
10348	69	Tg	I.mst	-	33	42	21	-	4	-	4	-	-	-	-	Q, P, k
10347	73	Tg	mst	2	33	52	-	-	13	-	-	*	13	-	-	Q, P
10346	75	Tg	mst	1	44	34	-	7	14	-	7	*	7	-	-	P, k
10345	77	Tg	mst	-	52	24	20	-	4	-	4	-	-	-	-	Q, P, k
10344	72	Tg	mst	2	37	52	-	2	7	-	-	*	7	-	-	Q, P
10343	76	Tg	mst	3	43	35	10	-	9	-	4	5	-	-	-	P
10342	74	Tg	mst	3	47	28	20	-	2	-	2	-	-	-	-	P
10341	68	Tg	mst	3	48	28	16	-	5	-	-	*	5	-	-	P
10340	70	Tg	mst	2	52	21	19	-	6	-	6	-	-	-	-	Q, P
10339	73	Tg	con	4	40	32	-	11	13	-	3	*	10	-	-	P
10338	75	Tg	con	2	35	43	-	7	17	-	5	*	8	-	-	P
10337	72	Tg	con	1	44	24	26	-	5	-	5	-	-	-	-	Q, P
10336	70	Tg	con	2	51	19	-	22	6	-	6	-	-	-	-	Q, P
10335	67	Mt	sst	-	40	32	24	-	4	-	4	-	-	-	-	P
10334	65	Mt	sst	4	37	32	21	-	6	-	-	*	6	-	-	P
10333	63	Mt	sst	-	40	31	24	-	5	-	5	-	-	-	-	P
10332	64	Mt	sst	-	52	23	22	-	3	-	3	-	-	-	-	P
10331	62	Mt	sst	-	47	29	21	-	3	-	3	-	-	-	-	P
10330	67	Mt	con	3	36	30	27	-	4	-	4	-	-	-	-	Q, P, k
10329	65	Mt	con	2	40	32	-	20	6	-	6	-	-	-	-	P
10328	64	Mt	con	-	62	5	25	-	8	-	8	-	-	-	-	Q, P

TABLE IV.1. (continued).

Sample	Local.	Strat.	Lith.	K	I	M	C		M/L	M/L Species						Others
							xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10327	50	Ur	sst	-	-	98	-	2	-	-	-	-	-	-	-	2P, An
10326	60	Ur	mst	2	56	18	20	-	4	-	4	-	-	-	-	P
10325	59	Ur	mst	-	44	32	21	-	3	-	3	-	-	-	-	P
10324	50	Ur	mst	-	60	18	-	17	5	-	5	-	-	-	-	Q, P
10323	46	Ur	mst	2	49	24	21	-	4	-	4	-	-	-	-	Q, P, k
10322	35	Ur	con	1	47	34	-	15	3	-	3	-	-	-	-	P
10321	50	Ur	con	-	59	24	14	-	6	-	6	-	-	-	-	P
10320	40	Ur	con	1	58	16	17	-	8	-	8	-	-	-	-	P, k
10319	59	Mt.M	sst	-	56	3	-	-	41	-	-	*	31	-	10	Q, P
10318	46	Mt.M	sst	2	61	10	-	22	5	-	5	-	-	-	-	Q, P
10317	42	Mt.M	sst	4	66	4	-	20	6	-	6	-	-	-	-	P
10316	57	Tw	mst	-	48	22	27	-	3	-	3	-	-	-	-	Q, P, k
10315	51	Tw	mst	2	51	21	23	-	3	-	3	-	-	-	-	Q, P, k
10314	57	Tw	con	2	50	20	19	-	9	-	5	*	4	-	-	Q, P, k
10313	51	Tw	con	3	64	5	-	20	8	-	8	-	-	-	-	Q, P
10312	39	Fy	sst	7	45	18	25	-	5							
10311	38	Fy	I.sst	4	69	2	-	18	7	-	7	-	-	-	-	P
10310	41	Fy	I.sst	4	38	39	-	13	6							
10309	46	Fy	I.sst	3	50	14	-	29	4	-	4	-	-	-	-	P
10308	43	Fy	I.sst	10	49	14	-	14	13	-	-	*	13	-	-	Q, P, k
10307	38	Fy	I.mst	4	65	6	-	21	4	-	4	-	-	-	-	P
10306	41	Fy	I.mst	3	24	58	-	7	8							
10305	56	Fy	I.mst	3	46	22	-	28	1	-	1	-	-	-	-	Q, P
10304	43	Fy	I.mst	7	48	31	-	11	3	-	3	-	-	-	-	Q, P, k
10303	32	Om	sst	-	74	-	-	26	-	-	-	-	-	-	-	P, Cs
10302	32	Om	I.sst	-	27	51	-	-	22	-	-	-	-	-	-	P, An, Am
10301	36	Om	I.sst	2	56	12	22	-	8	-	8	-	-	-	-	P
10300	32	Om	I.mst	-	43	44	-	6	7	-	-	*	7	-	-	P
10299	36	Om	I.mst	-	45	26	25	-	4	-	4	-	-	-	-	P
10298	36	Om	con	2	52	17	-	27	2	-	2	-	-	-	-	Q, P, k

TABLE IV .1. (continued).

Sample	Local.	Strat.	Lith.	K	I	M	C		M/L	M/L Species						Others
							xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10297	31	Pu	I.sst	-	5	94	-	1	-	-	-	-	-	-	-	2P, Am, Cs
10296	31	Pu	I.mst	-	50	32	-	12	6	-	-	*	6	-	-	Am, Cs
10295	38	Pu	sst	-	12	86	-	1	1	-	-	*	1	-	-	k, 2P, Am
10294	38	Pu	mst	-	50	5	-	14	31	-	7	*	24	-	-	P, Am
10293	38	Mg	c.sst	-	15	78	-	-	7	-	-	*	7	-	-	P
10292	25	Mg	c.sst	2	35	38	-	-	25	-	-	*	25	-	-	Q, P
10291	48	U.Mo	cgl	24	49	2	23	-	2	-	2	-	-	-	-	Q, P, k
10290	48	U.Mo	sst	10	54	12	10	-	14	-	4	*	10	-	-	Q, P, k
10289	31	U.Mo	c.sst	3	58	34	-	3	2	-	-	*	2	-	-	Q, P
10288	33	U.Mo	sst	5	45	23	-	14	13	-	-	*	13	-	-	P
10287	33	U.Mo	mst	3	50	20	-	7	20	-	-	*	20	-	-	P
10286	38	U.Mo	mst	9	57	3	20	-	11	-	2	*	9	-	-	P
10285	21	U.Mo	mst	6	50	18	13	-	13	-	-	*	13	-	-	Q
10284	44	U.Mo	mst	7	51	23	13	-	6	-	5	*	5	-	-	Q, P, k
10283	33	U.Mo	con	3	43	14	-	26	14	-	2	*	9	-	-	P
10282	88	My	mst	55	25	10	-	-	10	-	-	*	10	-	-	Q
10281	47	L.Mo	mst	9	23	41	-	-	27	-	-	*	27	-	-	Q, P, k
10280	33	L.Mo	sst	3	52	15	-	6	24	-	-	*	24	-	-	P, k
10279	33	L.Mo	sst	4	48	16	-	-	32	-	4	*	28	-	-	P
10278	16	L.Mo	sst	20	32	18	-	30	-	-	-	-	-	-	-	P
10277	21	L.Mo	sst	2	48	36	-	-	14	-	-	-	-	14	-	Q
10276	22	L.Mo	sst	7	35	56	-	-	2	-	-	*	2	-	-	Q, P
10275	54	L.Mo	sst	24	48	-	-	17	11	5	6	-	-	-	-	Q, P
10274	22	L.Mo	con	15	53	17	-	12	3	-	3	-	-	-	-	Q, P
10273	21	L.Mo	I.sst	7	79	9	-	-	5	-	-	*	5	-	-	P
10272	21	L.Mo	I.mst	4	34	48	-	3	11	-	-	*	11	-	-	Q, P

TABLE IV .1. (continued).

Sample	Local.	Strat.	Lith.	K	I	M	C		M/L	M/L Species						Others
							xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10271	27	Tmr	I.sst	10	33	43	-	-	14	-	-	*	14	-	-	Q, P
10270	45	Tmr	I.sst	8	48	23	-	-	21	-	-	*	21	-	-	-
10269	58	Tmr	I.sst	2	12	82	-	-	4			*	4			Q
10268	27	Tmr	I.mst	12	58	25	-	3	2	-	-	*	2	-	-	P
10267	45	Tmr	I.mst	18	54	22	-	3	3	-	3	-	-	-	-	P
10266	58	Tmr	I.mst	6	25	62	-	5	2	-	-	*	2	-	-	Q, P
10265	16	Tmt	c.mst	-	25	72	-	3	-	-	-	-	-	-	-	Q, P
10264	19	Tmt	c.mst	2	18	72	-	-	8	-	-	*	8	-	-	-
10263	20	Tmt	c.mst	3	20	67	-	7	3	-	-	*	3	-	-	Q, P
10262	21	Tmt	mst	5	27	54	-	7	7	-	-	*	7	-	-	Q, P
10261	30	Tmt	c.mst	3	32	41	-	12	12	-	-	*	12	-	-	P
10260	26	Tmt	c.mst	8	41	22	19	-	10	-	-	*	10	-	-	Q, P
10259	30	Tmt	con	3	33	36	-	8	20	-	-	*	20	-	-	P
10258	26	Tmt	con	8	46	9	-	33	4	-	4	-	-	-	-	Q, P
10257	30	Tmt	c.sst	3	31	43	-	8	15	-	-	*	15	-	-	P, Am
10256	30	Tmt	c.sst	3	48	22	11	11	4	-	4	-	-	-	-	Q, P, k
10255	30	Tmt	lst	2	26	67	-	5	7	-	-	-	-	-	-	-
10254	30	Tmt	lst	3	33	44	-	10	10	-	3	*	7	-	-	Q, P, k
10253	30	Tmt	lst	4	24	60	-	4	8	-	-	*	8	-	-	P
10252	34	Tmt	lst	10	36	37	-	-	17	-	-	-	17	-	-	Q, P
10251	34	Tmt	sty	5	18	57	-	7	13	-	-	*	13	-	-	Q, P, k
10250	15	Ot	lst	-	12	87	-	-	1	-	-	*	1	-	-	-
10249	15	Ot	lst	-	21	79	-	-	-	-	-	-	-	-	-	-
10248	15	Ot	lst	-	38	46	-	6	10	-	-	*	10	-	-	-
10247	28	Ot	lst	15	24	43	-	6	12	-	-	*	12	-	-	Q, P, k
10246	24	Ot	lst	1	43	54	-	2	-	-	-	-	-	-	-	-
10245	24	Ot	lst	6	42	47	-	3	2	-	-	*	2	-	-	-
10244	15	Ot	c.sst	-	33	61	-	-	6	-	-	*	6	-	-	P
10243	15	Wt	c.sst	-	20	75	-	-	5	-	-	*	5	-	-	P, Am

TABLE IV.1. (continued).

Sample	Local.	Strat.	Lith.	K	I	M	C		M/L	M/L Species						Others
							xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10242	15	Or	lst	-	15	85	-	-	-	-	-	-	-	-	-	Q, P
10241	28	Or	lst	3	5	87	-	-	5	-	5	30	-	-	10	P, k
10240	28	Or	lst	10	15	75	-	-	-	-	-	-	-	-	-	-
10239	23	Mp	c.sst	24	26	37	-	13	-	-	-	-	-	-	-	Q
10238	23	Mp	sst	18	31	32	-	-	19	-	-	*	19	-	-	Q
10237	3	T.A	mst	-	17	83	-	-	-	-	-	-	-	-	-	Q, k, Am, Cs
10236	4	T.A	mst	-	20	76	-	-	4	-	-	*	4	-	-	Q, P
10235	4	T.A	mst	-	10	88	-	-	2	-	-	*	2	-	-	Q, P
10234	4	T.A	mst	-	8	90	-	1	1	-	-	*	1	-	-	Q
10233	15	Ao	lst	-	39	61	-	-	-	-	-	-	-	-	-	P, k
10232	9	Ao	lst	-	12	87	-	-	1	-	-	*	1	-	-	Q, P, k, Am
10231	9	Ao	lst	-	22	76	-	-	2	-	-	*	2	-	-	Q, P, k, Am
10230	28	Ao	lst	25	23	49	-	3	-							
10229	15	Ao	c.sst	-	17	82	-	-	1	-	-	*	1	-	-	Am
10228	14	Ao	c.sst	-	12	86	-	1	1	-	-	*	1	-	-	Q, P, Am
10227	11	Ao	c.sst	-	11	89	-	-	-	-	-	-	-	-	-	Am, Cs
10226	28	Ao	c.sst	55	40	-	-	5	-	-	-	-	-	-	-	Q, P, k
10225	28	Ao	sst	3	77	14	-	-	6	-	6	-	-	-	-	Q
10224	28	Wh	lst	8	10	82	-	-	-	-	-	-	-	-	-	-
10223	4	Wh	c.sst	-	5	93	-	2	-	-	-	-	-	-	-	Q, P
10222	7	Wh	c.mst	-	12	87	-	1	-	-	-	-	-	-	-	Q, P, k, Am
10221	11	Wh	c.mst	-	7	91	-	2	-	-	-	-	-	-	-	Am, Cs
10220	4	Wh	c.mst	-	5	93	-	-	2	-	-	*	2	-	-	Q, P
10219	28	Wh	c.mst	-	6	93	-	-	1	-	-	*	1	-	-	Am
10218	5	G.M	lst	-	9	87	-	2	2	-	-	*	2	-	-	-
10217	5	G.M	c.mst	-	15	81	-	-	4	-	-	*	2	-	-	Q, P, k, Am
10216	5	G.M	c.sst	-	18	82	-	-	-	-	-	-	-	-	-	Q, P

TABLE IV .1. (continued).

Sample	Local.	Strat.	Lith.	K	I	M	C		M/L	M/L Species						Others
							xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10215	5	Mk	mst	22	15	35	-	-	28	-	-	*	28	-	-	Q
10214	2	Mk	mst	22	34	31	-	-	13	-	-	*	13	-	-	Q, k
10213	17	Mk	mst	77	18	1	-	1	3	-	2	*	1	-	-	Q
10212	17	Mk	con	73	22	-	-	2	3	-	-	*	3	-	-	Q
10211	4	Wk	mst	91	8	1	-	-	-	-	-	-	-	-	-	Q
10210	1	Wk	mst	89	6	-	-	-	5	-	-	-	-	-	-	Q
10209	28	Wk	mst	90	3	-	-	2	5	-	-	*	5	-	-	-
10208	28	Wk	mst	81	19	-	-	-	-	-	-	-	-	-	-	-
10207	18	Wk	mst	69	10	1	-	5	5	-	-	*	5	-	-	Q
10206	23	Wk	mst	76	6	-	-	4	14	-	-	-	14	-	-	Q
10205	23	Wk	sst	63	15	-	-	17	5	-	5	-	-	-	-	Q
10204	13	Js	mst	3	20	44	-	11	22	-	-	*	22	-	-	Q
10203	12	Js	mst	4	50	25	-	4	17	-	-	*	17	-	-	Q, P
10202	12	Js	con	1	-	98	-	-	17	-	-	*	1	-	-	P
10201	28	Ts	sst	-	23	1	-	31	45	-	5	30	-	-	10	P, k

RIVER AND HARBOUR SAMPLES

Sample	Strat.	K	I	M	C		M/L	M/L Species						Others
					xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10396	Rc	3	48	9	-	32	8	-	-	-	8	-	-	Q, P, k
10395	Rc	5	48	18	-	12	17	-	5	*	12	-	-	Q, P
10394	Rc	4	68	6	-	19	3	-	3	-	-	-	-	Q, P, k
10393	Rc	3	44	14	-	30	9	-	9	-	-	-	-	-
10392	Rc	3	48	13	20	-	16	-	16	-	-	-	-	-
10391	Rc	-	-	-	-	-	-	-	-	-	-	-	-	P
10390	Rc	-	-	-	-	-	-	-	-	-	-	-	-	P
10389	Rc	-	-	-	-	-	-	-	-	-	-	-	-	P, k
10388	Rc	5	60	6	-	26	3	1	2	-	-	-	-	Q, P
10387	Rc	2	50	14	-	27	7	2	5	-	-	-	-	Q, P

TABLE IV.1. (continued).

Sample	Strat.	K	I	M	C		M/L	M/L Species						Others
					xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
10386	Rc	3	48	10	24	-	15	-	7	-	8	-	-	Q, P
10385	Rc	3	40	20	16	-	21	-	13	*	8	-	-	Q, P
10384	Rc	4	44	7	10	-	35	-	6	11	18	-	-	Q, P, k
10383	Rc	13	59	6	-	10	12	-	-	-	9	-	-	P
10382	Rc	8	32	20	-	9	31	12	-	*	19	-	-	-
10381	Rc	5	37	24	-	7	27	5	-	*	22	-	-	Q, P
10380	Rc	10	35	42	-	6	7	-	-	*	7	-	-	Q, P, k
10379	Rc	11	30	40	-	5	14	-	3	*	11	-	-	Q
10378	Rc	10	25	55	-	2	8	-	-	*	8	-	-	Q
10377	Rc	12	26	55	-	2	5	-	-	*	5	-	-	Q
10376	Rc	4	54	10	-	8	24	-	-	*	9	-	15	Q, P
10375	Rc	9	36	18	-	3	34	-	-	*	13	-	21	P
10374	Rc	4	26	62	-	2	6	-	-	*	6	-	-	Q, P, Am
10373	Rc	4	15	66	-	7	8	-	-	*	8	-	-	-

WESTERN SHELF SAMPLES

Sample	Strat.	K	I	M	C		M/L	M/L Species						Others
					xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
A45	Rc	6	40	36	9	-	9	-	-	*	9	-	-	Q, P
A47	Rc	7	48	25	15	-	5	-	-	*	5	-	-	P
A51	Rc	7	50	27	8	-	8	-	-	*	8	-	-	P
B319	Rc	4	56	16	-	24	-	-	-	-	-	-	-	Q, P
B321	Rc	4	50	8	-	34	4	-	4	-	-	-	-	Q, P, k
B645	Rc	4	47	15	30	-	4	-	4	-	-	-	-	Q, k
B647	Rc	4	51	6	-	16	23	-	7	*	16	-	-	Q, P, k
B649	Rc	6	51	8	29	-	6	-	-	-	-	-	6	P
B675	Rc	4	43	37	5	-	11	-	-	*	11	-	-	Q, P
B676	Rc	8	53	6	-	26	7	3	-	-	-	-	4	Q, 2P
B682	Rc	4	48	26	16	-	6	-	-	*	6	-	-	Q, P
B778	Rc	9	21	67	-	1	2	-	-	*	2	-	-	Q, P
B780	Rc	15	24	60	-	-	1	-	-	*	1	-	-	Q, P
B781	Rc	8	26	56	-	-	10	-	-	*	10	-	-	P

TABLE IV.1. (continued).

Sample	Strat.	K	I	M	C		M/L	M/L Species						Others
					xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
B784	Rc	6	39	30	-	6	19	3	3	*	13	-	-	Q, P, k
B788	Rc	6	41	32	-	10	11	-	-	*	11	-	-	Q
B793	Rc	8	52	26	-	9	5	-	-	*	5	-	-	P
B795	Rc	4	40	40	-	10	6	-	-	*	6	-	-	Q
B796	Rc	4	41	16	-	11	28	5	-	*	23	-	-	Q, P
B804	Rc	6	45	20	-	20	9	9	-	-	-	-	-	Q, P
B816	Rc	5	44	12	-	7	32	4	7	*	21	-	-	Q, P, k
B818	Rc	4	50	9	24	-	13	3	2	*	8	-	-	Q
C169	Rc	4	50	8	18	-	20	-	-	*	10	10	-	Q, P, k
C262	Rc	8	56	5	-	17	14	-	-	*	8	6	-	P
C265	Rc	5	34	51	-	3	7	2	3	*	2	-	-	Q, k
C272	Rc	10	57	2	-	14	17	-	3	*	14	-	-	Q, P, k
C275	Rc	5	55	7	-	15	18	-	2	*	16	-	-	Q, P
C276	Rc	6	33	36	-	10	15	3	6	*	6	-	-	Q
C277	Rc	6	58	8	-	21	7	-	-	*	7	-	-	Q
C281	Rc	3	23	52	-	6	16	3	7	*	6	-	-	Q
C293	Rc	7	45	21	-	23	4	2	2	-	-	-	-	Q
C330	Rc	7	19	65	-	2	8	-	-	*	8	-	-	P
C332	Rc	13	44	29	-	4	10	-	-	*	10	-	-	Q, 2P
C340	Rc	5	16	71	-	2	6	-	-	*	6	-	-	Q
C348	Rc	10	50	11	-	15	14	-	6	*	8	-	-	Q
C355	Rc	9	50	11	-	18	12	-	3	*	9	-	-	-
C363	Rc	10	60	1	24	-	5	-	5	-	-	-	-	Q, P
C364	Rc	8	57	2	29	-	4	-	4	-	-	-	-	Q
C368	Rc	5	54	17	18	-	6	-	-	-	-	-	6	P
C372	Rc	4	44	22	21	-	9	-	2	-	-	-	-	P
C376	Rc	7	56	8	-	-	29	-	-	*	29	-	-	Q, k
C300	Rc	9	55	1	20	-	15	-	4	*	11	-	-	Q, P
C421	Rc	5	41	29	-	13	12	-	-	*	12	-	-	Q
C422.1	Rc	5	52	7	-	30	12	-	-	*	6	-	-	-
C422.4	Rc	5	47	10	-	30	8	-	-	*	8	-	-	—
C429.1	Rc	3	51	7	-	32	7	-	-	*	7	-	7	P
C429.4	Rc	3	63	3	-	27	4	-	-	*	4	-	4	P
C432	Rc	4	54	14	20	-	8	-	2	*	6	-	-	P
C434	Rc	4	53	13	22	-	8	-	2	*	6	-	-	P

TABLE IV.1. (continued).

Sample	Strat.	K	I	M			M/L	M/L Species						Others
					xC	hC		I-M	C-I	C-V	C-M	C-I-M	U.M/L	
C436	Rc	6	52	14	20	-	8	-	2	-	6	-	-	-
C438	Rc	6	60	14	16	-	4	-	-	*	4	-	-	Q, P, k
C450	Rc	4	58	5	-	16	17	-	-	*	17	-	-	Q, k

APPENDIX IV. CLAY FRACTION COMPOSITION

ONSHORE SAMPLES

Sample	Local.	Strat.	Lith.	I	C	M
10372	55	Rc	s	-	-	-
10371	61	Rc	s	-	-	-
10370	66	Rc	s	-	-	-
10369	78	Rc	s	0.11	-	0.20
10368	70	Rc	s	0.15	0.22	0.52
10367	21	Rc	s	0.19	-	0.14
10366	21	Rc	s	0.20	-	-
10365	8	Rc	s	0.09	-	-
10364	8	Rc	s	0.15	-	-
10363	86	Hw	sd	0.17	-	0.16
10362	86	Hw	sd	0.16	-	0.13
10361	85	Hw	sd	0.10	-	0.38
10360	81	Hw	sd	1.00	-	0.07
10359	86	Hw	md	0.86	-	0.05
10358	85	Hw	md	0.15	1.20	0.38
10357	81	Hw	md	0.06	0.22	0.60
10356	83	Wc	sst	0.08	0.10	0.17
10355	84	Wc	I.sst	0.05	1.00	0.49
10354	84	Wc	I.mst	0.05	0.79	0.41
10353	80	Wn	c.sst	0.04	0.05	0.28
10352	79	Wn	sst	0.05	1.00	0.42
10351	79	Wn	mst	0.04	0.06	0.18
10350	79	Wn	mst	0.05	0.06	0.38
10349	69	Tg	I.sst	0.04	0.10	0.39
10348	69	Tg	I.mst	0.09	0.13	0.64
10347	73	Tg	mst	0.07	0.09	0.66
10346	75	Tg	mst	0.04	0.08	0.52
10345	77	Tg	mst	0.02	0.03	0.29
10344	72	Tg	mst	0.10	0.27	0.52
10343	76	Tg	mst	0.04	0.05	0.50
10342	74	Tg	mst	0.03	0.04	0.01
10341	68	Tg	mst	0.04	0.05	0.46
10340	70	Tg	mst	0.03	0.04	0.04
10339	73	Tg	con	0.05	0.09	0.32
10338	75	Tg	con	0.07	0.10	0.51
10337	72	Tg	con	0.05	0.05	0.24
10336	70	Tg	con	0.04	0.07	0.26
10335	67	Mt	sst	0.05	0.04	0.51
10334	65	Mt	sst	0.05	0.03	0.35
10333	63	Mt	sst	0.03	0.03	0.53
10332	64	Mt	sst	0.03	0.03	0.42
10331	62	Mt	sst	0.02	0.03	0.42
10330	67	Mt	con	0.05	0.07	0.39
10329	65	Mt	con	0.06	0.04	0.48
10328	64	Mt	con	0.05	0.05	0.09

TABLE IV.2. Crystallinity of illite (I), chlorite (C) and montmorillonite (M) in the clay size (<2μ) fraction as determined by XRD analysis.

TABLE IV.2. (continued).

Sample	Local.	Strat.	Lith.	I	C	M
10327	50	Ur	sst	-	-	0.79
10326	60	Ur	mst	0.02	0.03	0.22
10325	59	Ur	mst	0.03	0.04	0.46
10324	50	Ur	mst	0.03	0.04	0.48
10323	46	Ur	mst	0.03	0.04	0.40
10322	59	Ur	con	0.05	0.06	0.38
10321	50	Ur	con	0.02	0.05	0.40
10320	46	Ur	con	0.03	0.09	0.33
10319	59	Mt.M	sst	0.08	0.16	0.10
10318	46	Mt.M	sst	0.05	0.10	-0.35
10317	42	Mt.M	sst	0.03	0.06	0.20
10316	57	Tw	mst	0.04	0.06	0.45
10315	51	Tw	mst	0.03	0.05	0.52
10314	57	Tw	con	0.04	0.06	0.32
10313	51	Tw	con	0.09	0.13	0.38
10312	39	Fy	sst	0.04	0.08	0.38
10311	38	Fy	I.sst	0.08	0.15	0.09
10310	41	Fy	I.sst	0.06	0.10	0.55
10309	56	Fy	I.sst	0.03	0.05	0.47
10308	43	Fy	I.sst	0.07	0.19	0.40
10307	38	Fy	I.mst	0.06	0.06	0.48
10306	41	Fy	I.mst	0.10	0.13	0.66
10305	56	Fy	I.mst	0.03	0.04	0.49
10304	43	Fy	I.mst	0.06	0.11	0.16
10303	32	Om	sst	0.53	1.00	-
10302	32	Om	I.sst	0.80	-	0.61
10301	36	Om	I.sst	0.02	0.02	0.35
10300	32	Om	I.mst	0.11	0.16	0.75
10299	36	Om	I.mst	0.04	0.06	0.64
10298	36	Om	con	0.04	0.05	0.54
10297	31	Pu	I.sst	1.00	1.00	0.65
10296	31	Pu	I.mst	0.12	0.17	0.58
10295	38	Pu	sst	0.39	1.00	0.93
10294	38	Pu	mst	0.09	0.37	0.31
10293	38	Mg	c.sst	0.25	0.20	0.83
10292	25	Mg	c.sst	0.15	0.18	0.60
10291	48	U.Mo	cgl	0.10	0.13	0.38
10290	48	U.Mo	sst	0.07	0.12	0.27
10289	31	U.Mo	c.sst	0.18	0.28	0.46
10288	33	U.Mo	sst	0.15	0.20	0.49
10287	33	U.Mo	mst	0.13	0.13	0.45
10286	38	U.Mo	mst	0.03	0.04	0.31
10285	21	U.Mo	mst	0.06	0.06	0.19
10284	44	U.Mo	mst	0.03	0.07	0.36
10283	33	U.Mo	con	0.14	0.16	0.34
10282	88	My	mst	0.20	-	0.26
10281	47	L.Mo	mst	0.10	-	0.23
10280	33	L.Mo	sst	0.09	0.25	0.34
10279	33	L.Mo	sst	0.04	0.15	0.45
10278	22	L.Mo	sst	0.05	0.04	0.50
10277	16	L.Mo	sst	0.08	-	0.26
10276	21	L.Mo	sst	0.20	0.31	0.19
10275	54	L.Mo	sst	0.05	0.13	-
10274	22	L.Mo	con	0.06	0.15	0.35
10273	21	L.Mo	I.sst	0.05	0.28	0.34
10272	21	L.Mo	I.mst	0.18	1.00	0.52

TABLE IV.2. (continued).

Sample	Local.	Strat.	Lith.	I	C	M
10271	27	Tmr	I.sst	0.17	0.16	0.73
10270	45	Tmr	I.sst	0.13	0.12	0.47
10269	58	Tmr	I.sst	0.40	0.58	0.94
10268	27	Tmr	I.mst	0.05	0.13	0.37
10267	45	Tmr	I.mst	0.05	0.09	0.28
10266	58	Tmr	I.mst	0.18	0.21	0.64
10265	16	Tmt	c.mst	0.16	0.38	0.68
10264	19	Tmt	c.mst	0.19	0.36	0.77
10263	20	Tmt	c.mst	0.14	0.29	0.68
10262	21	Tmt	mst	0.07	0.12	0.62
10261	30	Tmt	c.mst	0.15	0.13	0.77
10260	26	Tmt	c.mst	0.07	0.07	0.35
10259	30	Tmt	con	0.16	0.18	0.48
10258	26	Tmt	con	0.12	0.12	0.20
10257	30	Tmt	c.sst	0.14	0.11	0.72
10256	30	Tmt	c.sst	0.05	0.13	0.56
10255	30	Tmt	lst	0.41	0.75	0.58
10254	30	Tmt	lst	0.18	0.14	0.53
10253	30	Tmt	lst	0.15	0.19	0.70
10252	34	Tmt	lst	0.15	0.20	0.48
10251	34	Tmt	sty	0.22	0.10	0.48
10250	15	Ot	lst	0.34	0.67	0.57
10248	15	Ot	lst	0.57	-	0.20
10247	28	Ot	lst	0.39	0.18	0.47
10246	24	Ot	lst	0.18	0.63	0.38
10245	24	Ot	lst	0.18	0.55	0.40
10244	15	Ot	c.sst	0.27	0.50	0.46
10243	15	Wt	c.sst	0.37	0.67	0.66
10242	15	Or	lst	0.56	-	0.61
10241	28	Or	lst	-	0.67	0.81
10240	28	Or	lst	0.67	0.28	0.68
10239	23	Mp	c.sst	0.13	-	0.19
10238	23	Mp	sst	0.60	-	0.13
10237	3	T.A	mst	0.70	-	0.63
10236	4	T.A	mst	0.64	1.00	0.67
10235	4	T.A	mst	0.67	0.67	0.72
10234	4	T.A	mst	0.60	0.67	0.72
10233	15	Ao	lst	0.47	-	0.66
10232	9	Ao	lst	0.80	-	0.43
10231	9	Ao	lst	0.64	1.00	0.46
10230	28	Ao	lst	0.38	0.15	0.45
10229	15	Ao	c.sst	0.42	1.00	0.59
10228	14	Ao	c.sst	0.35	1.00	0.69
10227	11	Ao	c.sst	0.60	-	0.69
10226	28	Ao	c.sst	0.40	0.22	-
10225	28	Ao	sst	0.35	-	0.07
10224	28	Wh	lst	0.75	0.80	0.71
10223	4	Wh	c.sst	0.56	1.00	0.74
10222	7	Wh	c.mst	0.63	1.00	0.74
10221	11	Wh	c.mst	0.59	1.00	0.62
10220	4	Wh	c.mst	0.50	1.00	0.71
10219	28	Wh	c.mst	1.40	-	0.90

TABLE IV.2. (continued).

Sample	Local.	Strat.	Lith.	I	C	M
10218	5	G.M	lst	0.71	1.00	0.54
10217	5	G.M	c.mst	0.86	1.00	0.55
10216	5	G.M	c.sst	0.84	1.00	0.60
10215	5	Mk	mst	0.54	-	0.20
10214	2	Mk	mst	0.53	-	0.07
10213	17	Mk	mst	0.40	-	0.10
10212	17	Mk	con	0.45	-	-
10211	4	Wk	mst	1.09	-	0.17
10210	1	Wk	mst	1.00	-	-
10209	28	Wk	mst	1.43	-	-
10208	28	Wk	mst	0.70	-	-
10207	18	Wk	mst	0.59	-	0.10
10206	23	Wk	mst	0.88	-	-
10205	23	Wk	sst	0.39	-	-
10204	13	Js	mst	0.16	0.36	0.70
10203	12	Js	mst	0.08	0.10	0.07
10202	12	Js	con	-	1.20	0.72
10201	28	Ts	sst	0.13	0.27	0.50

RIVER AND HARBOUR SAMPLES

Sample	Local.	Strat.	I	C	M
10396	87	Rc	0.08	0.09	0.20
10395	82	Rc	0.09	0.09	0.57
10394	57	Rc	0.03	0.05	0.35
10393	58	Rc	0.13	0.09	0.42
10392	58	Rc	0.07	0.07	0.31
10391	55	Rc	-	-	-
10390	61	Rc	-	-	-
10389	66	Rc	-	-	-
10388	70	Rc	0.07	0.09	0.08
10387	52	Rc	0.07	0.06	0.37
10386	53	Rc	0.03	0.03	0.20
10385	49	Rc	0.05	0.04	0.36
10384	40	Rc	0.11	0.07	0.79
10383	37	Rc	0.09	0.07	0.37
10382	35	Rc	0.17	0.42	0.26
10381	29	Rc	0.12	0.12	0.32
10380	6	Rc	0.17	0.22	0.46
10379	5	Rc	0.55	0.85	0.32
10378	6	Rc	0.21	0.19	0.50
10377	6	Rc	0.23	0.37	0.49
10376	12	Rc	0.10	0.28	0.21
10375	12	Rc	0.31	0.33	0.43
10374	10	Rc	0.54	0.66	0.49
10373	10	Rc	0.66	0.72	0.27

TABLE IV.2. (continued).

WESTERN SHELF SAMPLES

Sample	Strat.	I	C	M
A45	Rc	0.10	0.17	0.30
A47	Rc	0.07	0.12	0.25
A51	Rc	0.07	0.07	0.55
B319	Rc	0.07	0.09	0.52
B321	Rc	0.05	0.06	0.20
B645	Rc	0.08	0.09	0.55
B647	Rc	0.01	0.08	0.80
B649	Rc	0.07	0.07	0.67
B675	Rc	0.05	0.08	0.28
B676	Rc	0.17	0.15	0.75
B682	Rc	0.08	0.08	0.32
B778	Rc	0.17	0.28	0.43
B780	Rc	0.19	0.47	0.58
B781	Rc	0.22	0.22	0.72
B784	Rc	0.12	0.20	0.41
B788	Rc	0.13	0.19	0.24
B793	Rc	0.06	0.07	0.32
B795	Rc	0.08	0.09	0.10
B796	Rc	0.16	0.11	0.72
B804	Rc	0.21	0.15	0.57
B816	Rc	0.08	0.09	0.69
B818	Rc	0.07	0.13	0.52
C169	Rc	0.08	0.08	0.66
C262	Rc	0.04	0.05	0.15
C265	Rc	0.12	0.19	0.38
C272	Rc	0.05	0.07	0.56
C275	Rc	0.03	0.04	0.44
C276	Rc	0.10	0.26	0.28
C277	Rc	0.07	0.09	0.43
C281	Rc	0.15	0.20	0.29
C293	Rc	0.13	0.14	0.31
C330	Rc	0.23	0.50	0.66
C332	Rc	0.09	0.14	0.53
C340	Rc	0.16	0.25	0.33
C348	Rc	0.12	0.10	0.16
C355	Rc	0.12	0.16	0.13
C363	Rc	0.02	0.02	0.75
C364	Rc	0.03	0.04	0.46
C368	Rc	0.05	0.06	0.27
C372	Rc	0.07	0.09	0.08
C376	Rc	0.20	0.31	0.64
C380	Rc	0.04	0.05	0.34
C421	Rc	0.14	0.13	0.29
C422.1	Rc	0.07	0.07	0.77
C422.4	Rc	0.09	0.09	0.53
C429.1	Rc	0.07	0.07	0.64
C429.4	Rc	0.05	0.08	0.60
C432	Rc	0.05	0.06	0.08
C434	Rc	0.05	0.06	0.23
C436	Rc	0.08	0.07	0.20
C438	Rc	0.05	0.06	0.39
C450	Rc	0.08	0.10	0.41

APPENDIX IV. CLAY FRACTION COMPOSITION

ONSHORE SAMPLES

		CLAY MINERAL ABUNDANCE (%)					CLAY MINERAL CRYSTALLINITY		
Strat.	Lith.	K	I	M	C	M/L	I	C	M
Hw	sd	1	45	46	–	8	0.36	–	0.15
Hw	mst	2	38	56	–	3	0.34	0.71	0.34
Wc	sst	6	44	37	7	6	0.08	0.10	0.17
Wc	I.sst	3	47	43	–	7	0.05	1.00	0.49
Wc	I.mst	5	49	44	–	2	0.05	0.79	0.41
Wn	sst	1	48	36	6	8	0.04	0.07	0.35
Wn	mst	2	43	32	17	6	0.04	0.06	0.28
Tg	I.sst	1	45	32	4	18	0.04	0.10	0.39
Tg	I.mst	–	33	42	21	4	0.09	0.13	0.64
Tg	mst	2	44	34	12	7	0.04	0.08	0.37
Mt	sst	1	43	29	22	4	0.04	0.03	0.45
Ur	sst	–	–	98	2	–	–	–	0.79
Ur	mst	1	52	23	20	4	0.03	0.04	0.39
Mt.M	sst	2	61	6	14	17	0.05	0.10	–0.02
Tw	mst	1	49	22	25	3	0.03	0.05	0.48
Fy	sst	7	45	18	25	5	0.04	0.08	0.38
Fy	I.sst	5	51	17	18	7	0.06	0.12	0.38
Fy	I.mst	4	46	29	17	4	0.06	0.08	0.45
Om	sst	–	74	–	26	–	0.53	1.00	–
Om	I.sst	1	41	32	11	15	0.41	0.50	0.48
Om	I.mst	–	44	35	15	5	0.07	0.11	0.69
Pu	sst	–	12	86	1	1	0.39	1.00	0.93
Pu	mst	–	50	5	14	31	0.09	0.37	0.31
Pu	I.sst	–	5	94	1	–	1.00	1.00	0.65
Pu	I.mst	–	50	32	12	6	0.12	0.17	0.58
Mg	c.sst	1	25	58	–	16	0.20	0.19	0.71
U.Mo	sst	6	52	23	9	10	0.13	0.20	0.41
U.Mo	mst	6	52	16	13	12	0.06	0.08	0.33
L.Mo	mst	9	23	41	–	27	0.10	–	0.23
L.Mo	sst	10	44	23	9	14	0.09	0.18	0.35
L.Mo	I.sst	7	79	9	–	5	0.05	0.28	0.34
L.Mo	I.mst	4	34	48	3	11	0.19	1.00	0.52
Tmr	I.sst	7	31	49	–	13	0.23	0.29	0.71
Tmr	I.mst	12	46	36	4	2	0.09	0.14	0.43
Tmt	mst	4	29	50	11	8	0.12	0.20	0.62
Tmt	sst	3	39	32	15	9	0.10	0.12	0.64
Tmt	lst	5	30	52	5	10	0.22	0.32	0.57

TABLE IV.3. Average clay fraction ($< 2\mu$) composition of onshore samples (formational average) and river and harbour samples (locality average) as determined by XRD analysis. K = kaolinite; I = illite; M = montmorillonite; C = total chlorite; xC = crystalline chlorite; hc = heat-labile chlorite; M/L = mixed-layer clay minerals.

TABLE IV.3. (continued).

Strat.	Lith.	CLAY MINERAL ABUNDANCE (%) K	I	M	C	M/L	CLAY MINERAL CRYSTALLINITY I	C	M
Ot	lst	4	30	59	3	4	0.33	0.51	0.40
Ot	sst	-	33	61	-	6	0.27	0.50	0.46
Wt	sst	-	20	75	-	5	0.37	0.67	0.66
Or	lst	4	12	82	-	2	0.61	0.61	0.70
Mp	sst	21	28	34	6	10	0.37	-	0.16
T.A	mst	-	14	84	-	2	0.70	0.78	0.69
Ao	lst	6	24	68	1	1	0.62	0.58	0.50
Ao	sst	12	31	54	1	2	0.42	0.74	0.51
Wh	lst	8	10	82	-	-	0.75	0.80	0.71
Wh	c.sst	-	5	93	2	-	0.56	1.00	0.74
Wh	c.mst	-	7	91	1	1	0.78	1.00	0.74
G.M	lst	-	9	87	2	2	0.71	1.00	0.54
G.M	c.mst	-	15	81	-	4	0.86	1.00	0.55
G.M	c. st	-	18	82	-	-	0.84	1.00	0.60
Mk	mst	40	22	22	-	15	0.49	-	0.12
Wk	mst	83	9	-	2	5	0.95	-	0.03
Wk	sst	63	15	-	17	5	0.39	-	-
Js	mst	4	35	34	8	19	0.12	0.23	0.39
Ts	sst	-	23	1	31	45	0.13	0.27	0.50

RIVER AND HARBOUR SAMPLES

Local.	Strat.	CLAY MINERAL ABUNDANCE (%) K	I	M	C	M/L	CLAY MINERAL CRYSTALLINITY I	C	M
87	Rc	3	48	9	32	8	0.08	0.09	0.20
82	Rc	6	46	25	8	15	0.09	0.09	0.57
57	Rc	4	69	6	19	3	0.03	0.05	0.35
59	Rc	3	44	14	30	9	0.10	0.08	0.36
55	Rc	-	-	-	-	-	-	-	-
61	Rc	-	-	-	-	-	-	-	-
65	Rc	-	-	-	-	-	-	-	-
70	Rc	5	60	6	26	3	0.07	0.09	0.08
52	Rc	2	50	14	27	7	0.07	0.06	0.37
53	Rc	3	48	10	24	15	0.03	0.03	0.20
49	Rc	3	40	20	16	21	0.05	0.04	0.36
40	Rc	4	44	7	10	35	0.11	0.07	0.79
37	Rc	13	59	6	10	12	0.09	0.07	0.37
35	Rc	8	32	20	9	31	0.17	0.42	0.26
29	Rc	5	37	24	7	27	0.12	0.12	0.32
6	Rc	11	29	48	4	8	0.29	0.40	0.44
12	Rc	7	45	14	5	28	0.20	0.30	0.32
10	Rc	4	20	64	4	7	0.60	0.69	0.38

APPENDIX V

CLAY MINERAL PARTICLE SIZE FRACTIONATION PROCEDURE

Particle size fractionation was carried out on 200 ml aliquots of dispersed mud size fraction suspensions (cf. Table 2.1) using a quadruple-head Multex MSE 20 centrifuge. Sedimentation times for the various size fraction separates required were calculated from Stokes' Law.

The size fractionation procedure is detailed in flow chart Table V.1 and the results are given in Table V.2.

For each size fractionated sample (Table V.2) the the <2μ size fraction was separated first to ensure that this fraction was separated in a manner identical to the <2μ size fraction of all other samples (Table IV.1). If for instance an aliquot of <1μ material had been removed prior to <2μ separation, the subsequently separated < 2μ fraction would have been slightly improverished in <1μ material.

Shaking between each size fractionation step ensured that the suspension remained in a dispersed state. Calgon (12 mg/l) was added if necessary to maintain dispersion. Between the separation of each size fraction samples received a further "wash" to remove any material of a size less than that required (cf. Table 2.1). Although further washes would produce a slightly better separation laboratory trials showed that on a time/benefit basis they were not necessary.

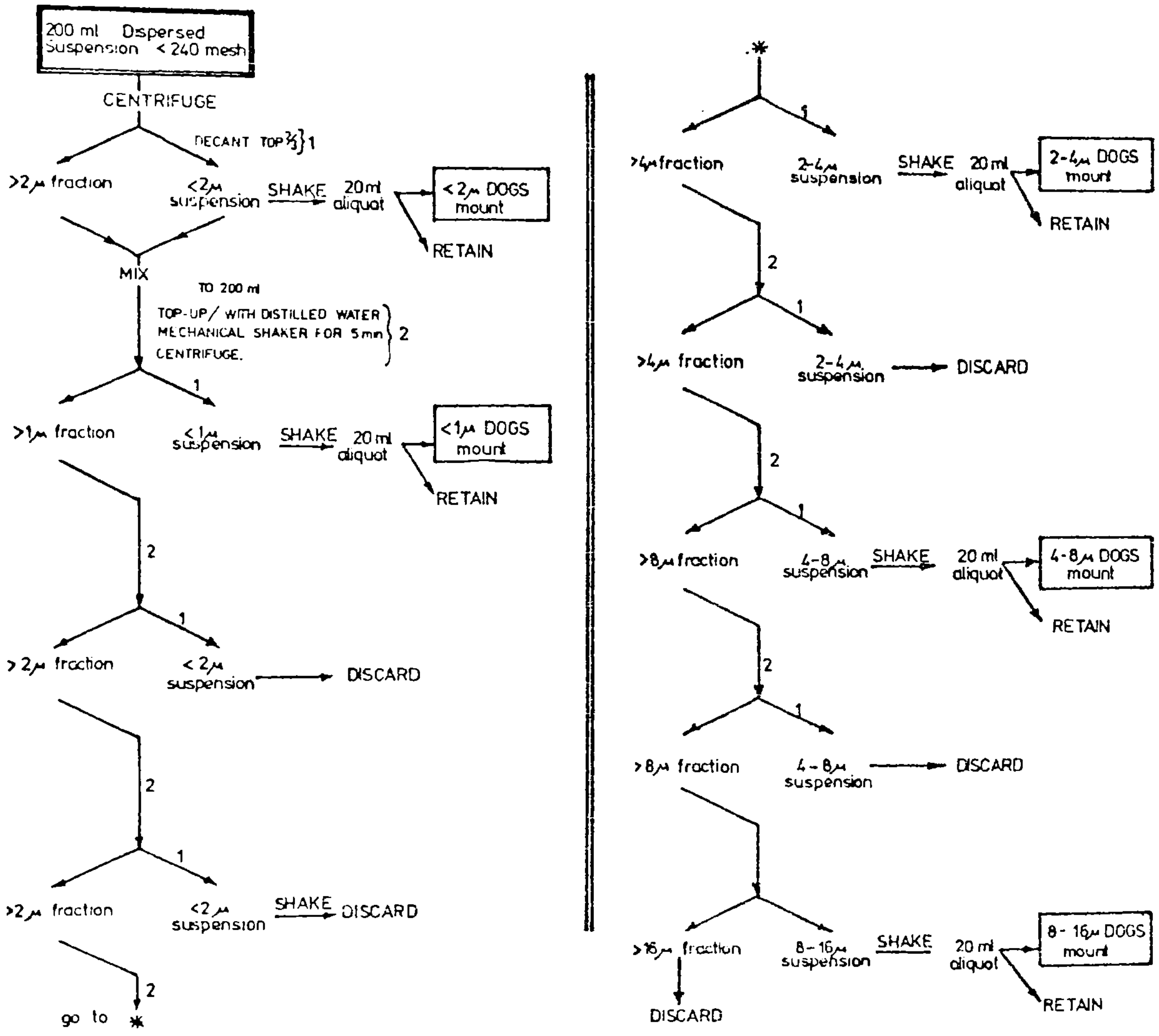


TABLE V.1. Flow chart summarising steps in clay mineral size fractionation procedure.

APPENDIX V.2. CLAY MINERAL DATA – SIZE FRACTIONATED SAMPLES

SAMPLE NUMBER	STRAT.	SIZE FRACTION (µ)	K	I	M	C		M/L	M/L Species				
			CLAY MINERAL ABUNDANCE (%)										
						xC	hC		I-M	C-M	C-V	C-I	U.M/L
8321	Rc	<1	4	37	21	–	13	25	–	13	*	12	–
"	"	<2	4	50	8	–	34	4	–	–	–	4	–
"	"	2–4	4	59	1	29	–	7	–	–	–	7	–
"	"	4–8	–	68	–	27	–	5	–	–	–	5	–
"	"	8–16	–	67	–	29	–	4	–	–	–	4	–
8778	Rc	<1	6	12	78	–	3	1	–	1	*	–	–
"	"	<2	9	21	67	–	1	2	–	2	*	–	–
"	"	2–4	5	50	19	21	–	5	–	–	–	5	–
"	"	4–8	–	65	5	27	–	3	–	–	–	3	–
"	"	8–16	–	74	–	26	–	–	–	–	–	–	–
10395	Rc	<1	6	46	25	–	8	15	–	9	*	6	–
"	"	<2	5	48	18	–	12	17	–	12	*	5	–
"	"	2–4	4	52	2	28	–	14	3	3	–	8	–
"	"	4–8	–	60	–	30	–	10	4	1	–	5	–
"	"	8–16	–	58	–	34	–	8	1	–	–	7	–
10389	Rc	<1	–	–	–	–	–	–	–	–	–	–	–
"	"	<2	–	–	–	–	–	–	–	–	–	–	–
"	"	2–4	–	–	–	–	–	–	–	–	–	–	–
"	"	4–8	–	–	–	–	–	–	–	–	–	–	–
"	"	8–16	–	–	–	–	–	–	–	–	–	–	–
10388	Rc	<1	5	52	10	–	27	6	2	–	–	4	–
"	"	<2	5	60	6	–	26	3	1	–	–	2	–
"	"	2–4	4	63	–	29	–	4	1	–	–	3	–
"	"	4–8	–	60	–	32	–	8	–	–	–	8	–
"	"	8–16	–	61	–	30	–	9	–	–	–	9	–
10383	Rc	<1	9	36	11	–	15	29	–	15	–	14	–
"	"	<2	13	59	6	–	10	12	–	9	–	–	–
"	"	2–4	7	57	3	22	–	11	–	5	–	2	–
"	"	4–8	–	56	–	32	–	12	3	–	–	9	–
"	"	8–16	–	49	–	37	–	14	–	14	–	–	–
10381	Rc	<1	5	33	28	–	6	28	6	22	*	–	–
"	"	<2	5	37	24	–	7	27	5	22	*	–	–
"	"	2–4	10	55	5	–	19	11	5	6	–	–	–
"	"	4–8	–	45	4	–	23	28	–	–	24	4	–
"	"	8–16	–	56	–	23	–	21	–	–	9	12	–
10380	Rc	<1	6	25	54	–	5	10	–	10	*	–	–
"	"	<2	10	35	42	–	6	7	–	7	*	–	–
"	"	2–4	7	44	30	–	10	9	–	4	*	5	–
"	"	4–8	–	42	24	–	22	12	–	7	*	5	–
"	"	8–16	–	59	15	18	–	8	–	8	–	–	–
10376	Rc	<1	5	34	8	–	12	41	–	24	*	–	17
"	"	<2	4	54	10	–	8	24	–	9	*	–	15
"	"	2–4	3	52	2	–	4	39	–	13	*	–	26
"	"	4–8	–	56	–	–	6	38	–	25	*	–	13
"	"	8–16	–	44	–	36	20	–	–	–	–	–	20
10374	Rc	<1	7	20	70	–	1	2	–	2	*	–	–
"	"	<2	4	26	62	–	2	6	–	6	*	–	–
"	"	2–4	3	59	17	–	10	11	–	11	*	–	–
"	"	4–8	–	45	13	–	7	35	–	7	*	–	28
"	"	8–16	–	40	–	–	30	30	–	–	–	–	30
10370	Rc	<1	–	–	–	–	–	–	–	–	–	–	–
"	"	<2	–	–	–	–	–	–	–	–	–	–	–
"	"	2–4	–	–	–	–	–	–	–	–	–	–	–
"	"	4–8	–	–	–	–	–	–	–	–	–	–	–
"	"	8–16	–	–	–	–	–	–	–	–	–	–	–

TABLE V.2. Clay mineral abundance (%) determined by XRD analysis for selected size fractions of 39 samples from the offshore and onshore study areas. K = kaolinite; I = illite; M = montmorillonite; C = total chlorite; xC = crystalline chlorite; hC = heat-labile chlorite; M/L = mixed-layer clay minerals.

TABLE V.2. (continued).

SAMPLE NUMBER	STRAT.	SIZE FRACTION (μ)	CLAY MINERAL ABUNDANCE (%) K	I	M	C xC	hC	M/L	M/L Species I-M	C-M	C-V	C-I	U.M/L
10368	Rc	<1	4	42	8	–	6	40	7	16	*	17	–
"	"	<2	3	36	5	–	9	47	5	30	*	12	–
"	"	2–4	4	36	2	–	13	45	8	14	*	23	–
"	"	4–8	–	44	–	35	–	21	8	–	–	13	–
"	"	8–16	–	46	–	26	–	28	10	–	–	18	–
10359	Hw	<1	–	–	100	–	–	–	–	–	–	–	–
"	"	<2	–	–	100	–	–	–	–	–	–	–	–
"	"	2–4	–	–	–	–	–	–	–	–	–	–	–
"	"	4–8	–	–	–	–	–	–	–	–	–	–	–
"	"	8–16	–	–	–	–	–	–	–	–	–	–	–
10354	Wc	<1	1	40	58	–	–	1	–	1	*	–	–
"	"	<2	5	50	43	–	–	2	–	2	*	–	–
"	"	2–4	2	72	–	–	7	19	–	–	11	8	–
"	"	4–8	–	77	–	9	–	14	–	–	7	7	–
"	"	8–16	–	80	–	9	–	11	–	–	6	5	–
10347	Tg	<1	2	34	51	–	–	13	–	13	*	–	–
"	"	<2	2	33	52	–	–	13	–	13	*	–	–
"	"	2–4	1	65	7	–	24	3	–	–	–	3	–
"	"	4–8	–	64	–	30	–	6	–	–	–	6	–
"	"	8–16	–	64	–	30	–	6	–	–	–	6	–
10341	Tg	<1	3	45	27	25	–	–	–	–	–	–	–
"	"	<2	3	49	27	16	–	5	–	5	*	–	–
"	"	2–4	2	55	10	30	–	3	–	–	–	3	–
"	"	4–8	–	62	–	32	–	6	–	–	–	6	–
"	"	8–16	–	65	–	29	–	6	–	–	–	6	–
10335	Mt	<1	1	33	43	12	–	11	–	11	*	–	–
"	"	<2	–	40	32	24	–	4	–	–	–	4	–
"	"	2–4	–	55	4	35	–	6	–	–	–	6	–
"	"	4–8	–	61	1	33	–	5	–	–	–	5	–
"	"	8–16	–	60	–	30	–	10	–	–	–	5	5
10334	Mt	<1	1	30	46	–	13	10	–	6	*	4	–
"	"	<2	4	37	32	21	–	6	–	–	–	6	–
"	"	2–4	3	53	2	36	–	6	–	–	–	6	–
"	"	4–8	–	58	–	35	–	7	–	–	–	7	–
"	"	8–16	–	56	–	28	–	16	–	–	–	9	7
10327	Ur	<1	–	–	97	–	3	–	–	–	–	–	–
"	"	<2	–	–	98	–	2	–	–	–	–	–	–
"	"	2–4	–	19	72	–	9	–	–	–	–	–	–
"	"	4–8	–	26	70	–	4	–	–	–	–	–	–
"	"	8–16	–	27	63	–	10	–	–	–	–	–	–
10324	Ur	<1	2	30	47	–	16	5	–	–	–	5	–
"	"	<2	–	60	18	–	17	5	–	–	–	5	–
"	"	2–4	–	73	1	16	–	10	–	–	–	10	–
"	"	4–8	–	70	–	22	–	8	–	–	–	8	–
"	"	8–16	–	71	–	22	–	7	–	–	–	7	–
10323	Ur	<1	1	52	31	–	16	–	–	–	–	–	–
"	"	<2	2	49	24	21	–	4	–	–	–	4	–
"	"	2–4	–	65	5	25	–	5	–	–	–	5	–
"	"	4–8	–	66	1	25	–	8	–	–	–	8	–
"	"	8–16	–	69	1	23	–	7	–	–	–	7	–
10317	Mt.M	<1	6	58	20	–	4	12	–	–	5	7	–
"	"	<2	4	66	4	–	20	6	–	–	–	6	–
"	"	2–4	1	69	2	21	–	7	–	–	–	7	–
"	"	4–8	–	68	–	26	–	7	–	–	–	7	–
"	"	8–16	–	68	–	27	–	5	–	–	–	5	–
10311	Fy	<1	5	68	5	–	8	14	–	14	*	–	–
"	"	<2	4	69	2	–	18	7	–	–	–	7	–
"	"	2–4	3	63	–	–	26	8	–	–	–	8	–
"	"	4–8	–	62	–	31	–	7	–	–	–	7	–
"	"	8–16	–	59	–	35	–	6	–	–	–	6	–

TABLE V.2. (continued).

SAMPLE NUMBER	STRAT.	SIZE FRACTION (μ)	CLAY MINERAL ABUNDANCE (%)										
			K	I	M	C		M/L	M/L Species				
						xC	hC		I-M	C-M	C-V	C-I	U.M/L
10307	Fy	< 1	1	43	30	–	22	4	–	4	*	–	–
"	"	< 2	4	66	5	–	21	4	–	–	–	4	–
"	"	2–4	2	66	1	24	–	7	–	–	–	7	–
"	"	4–8	–	58	1	34	–	7	–	–	–	7	–
"	"	8–16	–	58	–	33	–	9	–	–	–	9	–
10309	Fy	< 1	4	39	42	–	10	5	–	5	*	–	–
"	"	< 2	3	49	12	–	24	12	–	8	*	4	–
"	"	2–4	–	55	5	32	–	8	–	–	–	8	–
"	"	4–8	–	57	2	35	–	6	–	–	–	6	–
"	"	8–16	–	55	1	38	–	6	–	–	–	6	–
10305	Fy	< 1	4	51	23	–	16	6	–	4	*	2	–
"	"	< 2	3	49	22	25	–	1	–	–	–	1	–
"	"	2–4	2	57	4	29	–	8	–	–	–	8	–
"	"	4–8	–	57	2	32	–	9	–	–	–	9	–
"	"	8–16	–	63	2	29	–	6	–	–	–	6	–
10297	Pu	< 1	–	5	95	–	–	–	–	–	–	–	–
"	"	< 2	–	5	95	–	–	–	–	–	–	–	–
"	"	2–4	–	7	93	–	–	–	–	–	–	–	–
"	"	4–8	–	32	68	–	–	–	–	–	–	–	–
"	"	8–16	–	64	36	–	–	–	–	–	–	–	–
10296	Pu	< 1	3	55	31	–	7	4	–	4	*	–	–
"	"	< 2	–	49	32	–	12	7	–	7	*	–	–
"	"	2–4	2	61	20	–	7	10	–	6	*	–	–
"	"	4–8	–	60	7	18	–	15	–	10	–	5	–
"	"	8–16	–	62	7	27	–	4	–	–	–	4	–
10293	Mg	< 1	–	16	78	–	1	5	–	5	*	–	–
"	"	< 2	–	14	79	–	–	7	–	7	*	–	–
"	"	2–4	–	31	59	7	–	3	–	3	*	–	–
"	"	4–8	–	48	30	15	–	7	–	4	*	3	–
"	"	8–16	–	51	20	24	–	5	–	–	–	5	–
10285	U.Mo	< 1	4	38	46	–	–	12	–	12	*	–	–
"	"	< 2	6	50	18	13	–	13	–	13	*	–	–
"	"	2–4	7	58	3	10	–	22	2	–	20	–	–
"	"	4–8	–	61	2	23	–	14	3	–	11	–	–
"	"	8–16	–	65	–	29	–	6	6	–	–	–	–
10280	L.Mo	< 1	3	49	11	–	7	30	–	–	22	–	8
"	"	< 2	3	53	14	–	6	24	–	–	24	–	–
"	"	2–4	3	57	3	–	6	31	–	–	20	–	11
"	"	4–8	–	68	–	–	10	22	–	–	22	–	–
"	"	8–16	–	69	–	–	25	6	–	–	6	–	–
10270	Tmr	< 1	9	46	27	–	–	18	–	18	*	–	–
"	"	< 2	8	48	23	–	7	14	–	14	*	–	–
"	"	2–4	11	53	5	–	25	6	–	6	*	–	–
"	"	4–8	–	61	1	30	–	8	–	–	8	–	–
"	"	8–16	–	57	–	34	–	9	–	–	9	–	–
10267	Tmr	< 1	8	44	16	18	–	14	–	14	–	–	–
"	"	< 2	18	54	22	3	–	3	–	–	–	3	–
"	"	2–4	11	49	17	–	20	3	–	–	–	3	–
"	"	4–8	–	60	5	–	32	3	–	–	–	3	–
"	"	8–16	–	58	2	40	–	–	–	–	–	–	–
10265	Tmr	< 1	–	14	81	–	–	5	–	5	*	–	–
"	"	< 2	–	24	73	–	3	–	–	–	–	–	–
"	"	2–4	–	33	54	–	13	–	–	–	–	–	–
"	"	4–8	–	32	57	–	11	–	–	–	–	–	–
"	"	8–16	–	34	52	–	14	–	–	–	–	–	–
10250	Ot	< 1	–	17	80	–	2	1	–	1	*	–	–
"	"	< 2	–	12	87	–	–	1	–	1	*	–	–
"	"	2–4	–	55	34	–	–	11	–	6	*	–	5
"	"	4–8	–	35	45	–	–	20	–	13	*	–	7
"	"	8–16	–	63	17	–	–	20	–	12	*	–	8

TABLE V.2. (continued).

SAMPLE NUMBER	STRAT.	SIZE FRACTION (μ)	CLAY MINERAL ABUNDANCE (%)										
			K	I	M	C		M/L	M/L Species				
						xC	hC		I-M	C-M	C-V	C-I	U.M/L
10234	T.A	<1	-	7	91	-	-	2	-	2	*	-	-
"	"	<2	-	8	90	-	1	1	-	1	-	-	-
"	"	2-4	-	23	74	-	-	3	-	3	*	-	-
"	"	4-8	-	24	67	-	-	9	-	9	*	-	-
"	"	8-16	-	32	47	-	7	14	-	14	*	-	-
10219	Wh	<1	-	4	95	-	-	1	-	1	*	-	-
"	"	<2	-	6	93	-	-	1	-	1	*	-	-
"	"	2-4	-	18	76	-	-	6	-	6	*	-	-
"	"	4-8	-	20	71	-	3	6	-	6	*	-	-
"	"	8-16	-	30	66	-	3	1	-	1	*	-	-
10215	Mk	<1	20	10	56	-	-	14	-	14	*	-	-
"	"	<2	22	15	35	-	-	28	-	28	*	-	-
"	"	2-4	20	36	28	-	-	16	-	16	*	-	-
"	"	4-8	25	41	11	-	3	20	-	4	*	-	-
"	"	8-16	15	43	13	-	4	25	-	5	*	-	-
10201	Wk	<1	-	26	5	-	23	46	-	-	23	-	23
"	"	<2	-	23	1	-	30	46	-	-	31	5	10
"	"	2-4	-	30	-	-	28	42	-	14	*	8	20
"	"	4-8	-	30	-	-	25	45	-	25	*	10	10
"	"	8-16	-	50	-	-	25	25	-	25	*	-	-

APPENDIX VI

CLAY MINERAL DATA - DOGS VS SOGS

Statistical analysis was undertaken on the clay mineral data (Appendix Table VI.1) to compare the results of the smear-on-glass slide (SOGS) and the dropper-on-glass slide (DOGS) techniques. Firstly a one way analysis of variance was carried out to see if the results of each method differed significantly from one another.

1. For DOGS I (I - illite) versus SOGS I two hypotheses were tested:

H_o - There is no difference in the two methods.
H_I - This is a difference in the two methods.

Test Statistic $F = \frac{\text{Ms regression}}{\text{Ms residual}}$

Source of Variation	Degrees of Freedom	Sum of Squares	Mean of Squares (Ms)
Total		85863	
By Regression	1	83873	83973
Residual (error)	47	1989.3	42.331
Correlation Coefficient R	0.98835		
95% Limits on R	0.97919-0.99349		
F - Ratio Statistic (calc.)	1981.4		

Critical Value of Test Statistic

$F_{0.01}$ (1,47) = 7.2

Rejection Region

F Ratio (calc.) Critical Value

Conclusion

Since the computed value of F exceeds $F_{0.01}$ i.e. 1981.4 > 7.2 reject Ho (null hypothesis) that there is no difference and accept the alternative that there is a significant difference between the two methods at the 99% confidence level.

A simple linear correlation between the two methods showed that

$$R > 0.9$$

i.e. the DOGS I versus SOGS I relationship is strongly linear.

2. For DOGS M (M = Montmorillonite) versus SOGS M two hypotheses were tested:

H_o - There is no difference in the two methods.
H_I - There is a difference in the two methods.

$$\text{Test Statistic } F = \frac{\text{Ms regression}}{\text{Ms residual}}$$

Source of Variation	Degrees of Freedom	Sum of Squares	Mean of Squares (Ms)
Total		100600	
By Regression	1	98217	98217
Residual (error)	47	2381.5	50.67
Correlation Coefficient R	0.98809		
95% Limits on R	0.97874-0.99334		
F - Ratio Statistic (calc.)	1938.4		

Critical Value of Test Statistic

$$F_{0.01} \quad (1,47) = 7.2$$

Rejection Region

F Ratio (calc.) > Critical Value

Conclusion

Since the computed value of F exceeds $F_{0.01}$ i.e. 1938.4 > 7.2 reject Ho (null hypothesis) that there is no difference and accept the alternative that there is a significant difference between the two methods at the 99% confidence level.

A simple linear correlation between the two methods showed that

$$R > 0.9$$

i.e. the DOGS M versus SOGS M relationship is strongly linear.

3. For DOGS CKM (CKM = chlorite + kaolinite + mixed layer clays) versus SOGS CKM two hypotheses were tested:

Ho - There is no difference in the two methods.
H$_I$ - There is a difference in the two methods.

$$\text{Test Statistic } F = \frac{\text{Ms regression}}{\text{Ms residual}}$$

Source of Variation	Degrees of Freedom	Sum of Squares	Mean of Squares (Ms)
Total		50209	
By Regression	1	48423	48423
Residual (error)	47	1786.4	38.008
Correlation Coefficient R	0.98205		
95% Limits on R	0.96803-0.98995		
F = Ratio Statistic (calc.)	1274		

Critical Value of Test Statistic

$$F_{0.01}\ (1,47) = 7.2$$

Rejection Region

F Ratio (calc.) > Critical Value

Conclusion

Since the computed value of F exceeds $F_{0.01}$ i.e. 1274 > 7.2 reject Ho (null hypothesis) that there is no difference and accept the alternative that there is a significant difference between the two methods at the 99% confidence level.

A simple linear correlation between the two methods showed that

$$R > 0.9$$

i.e. the DOGS CKM versus SOGS CKM relationship is strongly linear.

APPENDIX VI. CLAY MINERAL DATA – DOGS vs. SOGS

SAMPLE NUMBER	STRAT.	SAMPLE PREPARATION METHOD	CLAY MINERAL ABUNDANCE (%)		
			I	M	C + K + M/L
8319	Rc	D	56	16	28
"	"	S	48	20	30
8321	Rc	D	50	8	39
"	"	S	50	8	39
8778	Rc	D	21	67	12
"	"	S	27	53	20
8780	Rc	D	24	60	16
"	"	S	27	54	19
8784	Rc	D	39	30	31
"	"	S	37	37	26
8788	Rc	D	41	32	27
"	"	S	36	39	25
8795	Rc	D	40	40	20
"	"	S	44	23	33
8796	Rc	D	41	16	43
"	"	S	46	22	32
C421	Rc	D	41	29	30
"	"	S	45	24	31
C429	Rc	D	56	11	33
"	"	S	51	15	34
C434	Rc	D	53	13	34
"	"	S	50	18	32
C450	Rc	D	58	5	37
"	"	S	51	14	35
10395	Rc	D	48	18	34
"	"	S	47	16	32
10393	Rc	D	44	14	42
"	"	S	47	11	37
10388	Rc	D	60	6	34
"	"	S	51	11	36
10386	Rc	D	48	10	42
"	"	S	45	11	33
10384	Rc	D	44	7	49
"	"	S	49	10	33
10380	Rc	D	35	42	23
"	"	S	27	56	17
10381	Rc	D	37	24	39
"	"	S	42	16	39
10375	Rc	D	36	18	46
"	"	S	44	16	40
10374	Rc	D	26	62	12
"	"	S	30	60	10
10255	Wc	D	47	43	10
"	"	S	62	34	4
10254	Wc	D	49	44	7
"	"	S	59	34	7
10341	Tg	D	48	28	24
"	"	S	48	18	32
10334	Tg	D	37	32	31
"	"	S	49	17	33

TABLE VI.1. Comparison of clay mineral abundance % determined by XRD analysis for the <2μ size fraction of 48 selected samples prepared by both the dropper-on-glass slide (D) and the smear-on-glass slide (S) techniques. I = illite; M = montmorillonite; C + K + M/L = chlorite plus kaolinite plus mixed-layer clay minerals.

TABLE VI.1.(continued).

SAMPLE NUMBER	STRAT.	SAMPLE PREPARATION METHOD	CLAY MINERAL ABUNDANCE (%)		
			I	M	C + K + M/L
10327	Ur	D	–	98	2
"	"	S	–	95	5
10324	Ur	D	60	18	22
"	"	S	56	13	29
10317	Mt.M	D	66	20	14
"	"	S	57	5	34
10311	Fy	D	69	2	29
"	"	S	58	5	32
10307	Fy	D	65	6	29
"	"	S	55	7	37
10297	Pu	D	5	94	1
"	"	S	11	85	4
10296	Pu	D	50	32	18
"	"	S	57	24	19
10293	Mg	D	15	78	7
"	"	S	22	71	7
10287	U.Mo	D	50	20	30
"	"	S	50	16	34
10280	L.Mo	D	52	15	33
"	"	S	54	10	36
10270	Tmr	D	48	23	29
"	"	S	51	19	30
10267	Tmr	D	54	22	24
"	"	S	53	19	28
10256	Tmt	D	48	22	30
"	"	S	47	18	35
10250	Ot	D	12	87	1
"	"	S	24	73	3
10243	Wt	D	20	75	5
"	"	S	23	76	1
10242	Or	D	15	85	–
"	"	S	27	73	–
10234	T.A	D	8	90	2
"	"	S	18	76	6
10227	Ao	D	11	89	–
"	"	S	17	82	–
10222	Wh	D	12	87	1
"	"	S	13	85	2
10215	Mk	D	15	35	50
"	"	S	19	31	50
10210	Wk	D	6	–	94
"	"	S	3	–	97
10203	Js	D	50	25	25
"	"	S	52	13	29
10201	Ts	D	23	1	76
"	"	S	31	–	69

APPENDIX VII

SAMPLE NUMBER	SAMPLE TYPE	TEM	SEM
B 645	W.Sd	x	
B 778	"		x
B 780	"	x	
B 796	"	x	
C 275	"	x	
C 363	"	x	
C 434	"		x
N 377	W.St		x
N 378	"		x
N 379	"		x
N 382	"		x
N 383	"		x
N 384	"		x
10388	Rc	x	
10376	"	x	
10359	H	x	
10354	W	x	x
10347	Tg	x	
10341	"		x
10334	Mt	x	x
10327	Ur	x	x
10324	"	x	x
10320	Mt.M		x
10317	"	x	x
10308	Tw	x	
10311	Fy		x
10307	Fy		x
10301	Om	x	
10299	"	x	
10297	Pu		x
10296	"	x	x
10290	U.Mo	x	
10287	"	x	
10286	"		x
10280	L.Mo		x
10279	"	x	
10270	Tmr	x	
10267	"	x	
10256	Tmt	x	x
10236	T.A	x	
10227	Ao	x	
10222	Wh		x
10219	"	x	
10215	Mk		x
10213	"	x	
10212	"	x	
10210	Wk		x
10208	"	x	
10203	Js	x	x
10202	Js		x
10201	Ts	x	x

TABLE VII.1. Samples examined by transmission electron microscopy (TEM) and scanning electron microscopy (SEM). Sample type abbreviations defined in Table 1.1. W.Sd = Western shelf surficial sediment. W.St = Western shelf seston.

APPENDIX VIII

X-RAY FLUORESCENCE SPECTROMETRY

SAMPLE PREPARATION AND ANALYSIS

The sediment powders analysed were those prepared for XRD modal examination which had been pre-treated with both hydrogen peroxide and acetic acid. The removal of atomic carbon (present in organic matter and $CaCO_3$) from the samples minimised matrix effects and ensured that all the calcium left in the samples was that bonded within the lattice of clay minerals, feldspars and other silicate minerals. The samples, thoroughly mixed by the milling process were then split by pouring into a cone and quartering. Powder samples were mounted as loose powders in Ortec powder sample holders.

The major problem associated with the elemental analysis of discrete clay minerals is isolating them from multicomponent mixtures. The separation of chlorite, illite and mixed-layer clays is very difficult and according to Dunoyer de Segonzac (1970, p.326) "...mixed-layer C-M, corrensite, C-V and chlorite cannot be isolated for chemical analysis." In selected samples from this study both kaolinite and montmorillonite were isolated by segregating a size fraction in which kaolinite or montmorillonite was virtually the only clay mineral present. The size fractions were separated by repeated centrifugation and washing, dried at 40°C, ground to a fine powder, and mounted in the Ortec powder sample holders for analysis. Rock and clay mineral standards were crushed to fine powders and mounted in the same manner.

Analysis was carried out on an Ortec XRF multichannel analyser at the following settings:

MA 50
KV 20 Output voltage 1/50
Tube Mo
Count time 200 sec
Energy scale 40.96/1 Kev Energy calibration 7/34
Preset controls Base 2
Multiplier 100
Analysis region Lower level 0/20
Upper level 10/50

The following elements were determined: Al, Si, K, Ca, Ti, V, Cr, Mn, and Fe. Although the presence or absence of Na and Mg could be detected, machine calibration at the time of analysis was not sufficiently sophisticated to permit their quantitative determination.

MACHINE CALIBRATION AND INTERPRETATION OF DATA

To allow quantitative determination of element concentration in the offshore and onshore samples of rock and clay standards of known composition (Table VIII.1) were

	Dunite	Diabase	Syenite	Basalt	Clay	Granite	Limestone	Montmorillonite	Montmorillonite
	USGS-DTS-1	USGS-W-1	SSC-SY-1					API-23	API-27
Si	18.90	24.57	28.40	21.60	31.40	33.94	8.48	23.33	27.36
Ti	0.08	0.68	0.06	1.40	0.17	0.17	0.13	0.14	0.07
Al	0.13	7.76	6.00	7.06	12.20	7.45	3.00	9.10	10.38
Fe	6.47	7.12	3.82	9.13	0.16	1.30	1.20	1.76	2.26
Mn	-	0.13	0.20	0.10	-	-	0.03	0.03	-
Ca	0.11	7.42	4.71	6.61	0.05	0.94	26.90	1.65	0.18
K	-	0.70	3.53	1.46	0.12	5.60	0.96	0.23	0.26

TABLE VIII.1. Elemental composition (weight %) of geochemical standards.

analysed under the conditions described in the previous section, the peak height intensities above background for elements in the standard powders were plotted against elemental concentrations and the derived intensity-concentration curves were used to determine the elemental concentrations in "unknown" sediments and clays. The relative amounts of V and Cr in all samples were determined by comparing the reflected intensities for each of these elements.

The diverse range of standards used places limits on the interpretation of the quantitative data. In theory, standards should be as close as possible in composition to the "unknowns" and preferably should be rocks and minerals from the materials under analysis. However, the lack of suitable standards at this early stage of machine calibration necessitated the use of those listed above. Despite their large range in composition statistical analysis showed good linear relationships existed between element concentration and reflected intensity for Si, Ca, Ti, and Fe. (Fig. VIII.1). A linear relationship existed for K at low concentration values but a parabolic line was a better fit at higher concentrations (Fig. VIII.1). The linear correlation coefficient derived for the intensity-concentration line of each mineral is given in Figure VIII.1.

Chemical analyses of kaolinite and montmorillonite material isolated from selected samples is presented in Table VIII.2. The montmorillonite analyses suggest that the dioctahedral Whaingaroa and Taumatamaire montmorillonites belongs to the "montmorillonite-beidellite" series, and are probably of the montmorillonite type (Weaver and Pollard,

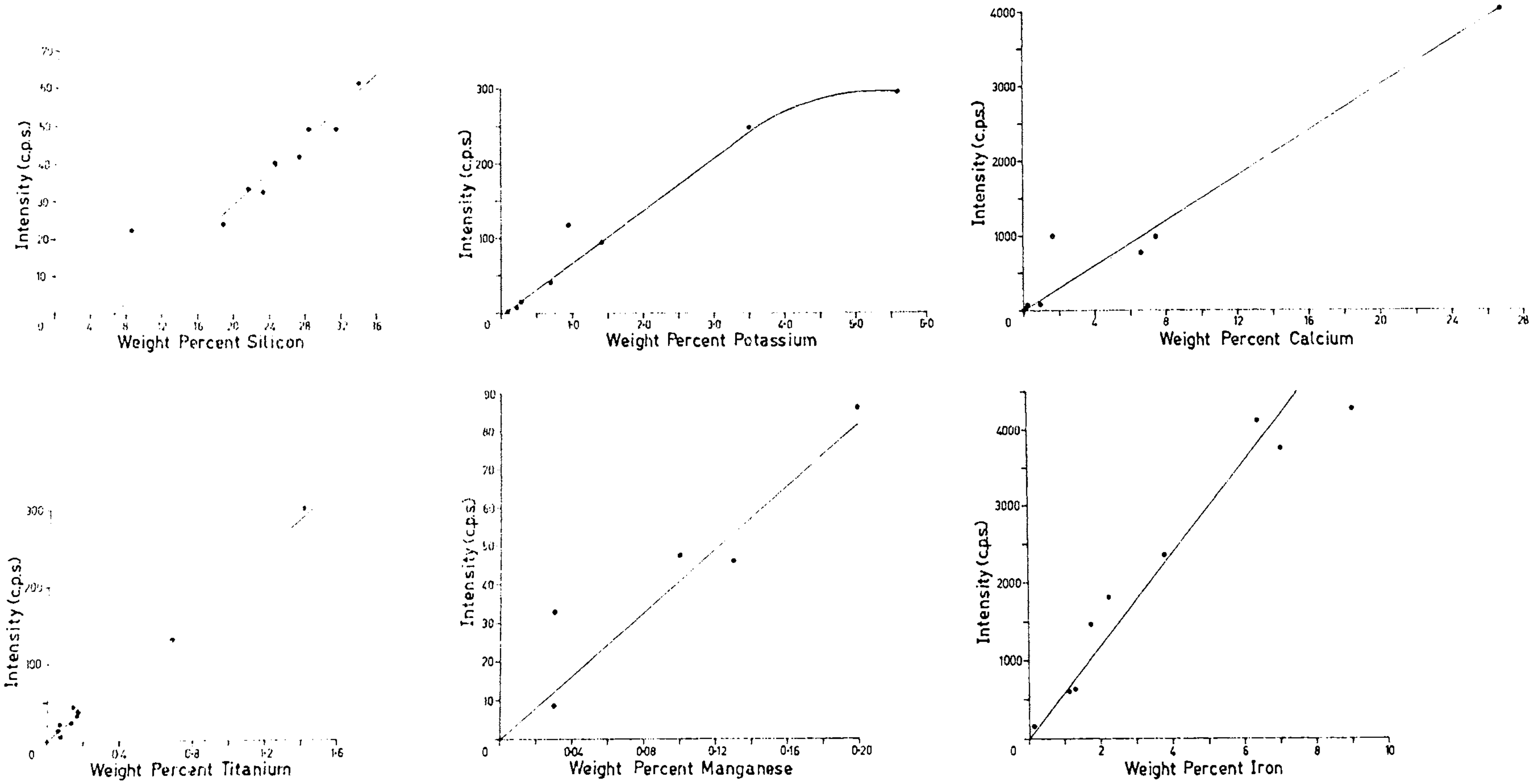


Fig. VIII.1. Intensity-concentration curves for Si, K, Ca, Ti, Mn and Fe derived from XRF analysis of rock and clay mineral standards. The linear correlation coefficients for the intensity-concentration plots of each mineral are as follows: Si = 0.993; K = 0.981; Ca = 0.979; Ti = 0.995; Mn = 0.824 and Fe = 0.987.

1973, pp.55-86). The trioctahedral Purupuru and Urenui montmorillonites (both of volcanic origin) belong to the "stevensite-saponite" series, the low Al_2O_3 content being suggestive of stevensite (Weaver and Pollard, 1973, pp.55-86).

	Montmorillonite				Kaolinite		
	1	2^+	3	4	5	6	7
SiO_2	62.34	59.75	27.43	26.84	45.60	46.72	48.75
TiO_2	1.37	1.17	1.50	0.80	2.17	1.90	1.72
Al_2O_3	18.70	19.12	-	-	41.98	43.92	36.56
$Fe_2O_3^*$	3.86	8.31	7.29	8.02	2.31	3.26	9.34
MnO	0.17	0.04	0.20	0.02	-	0.00	0.04
MgO^x	1.93	2.87	2.63	2.26	n.d.	n.d.	n.d.
CaO	1.98	1.46	3.56	0.77	0.48	0.12	0.46
K_2O	1.68	3.00	0.89	0.26	0.96	2.24	4.86

* represents total iron as $FeO + Fe_2O_3$

\+ some illite and chlorite present

x determined by atomic absorption spectrometry

n.d. not determined

1 Whaingaroa siltstone, sample 10219

2 Taumatamaire mudstone, sample 10265

3 Purupuru sandstone, sample 10297

4 Urenui sandstone, sample 10237

5 Waikato Coal Measures mudstone, sample 10210

6 Waikato Coal Measures mudstone, sample 10206

7 Waikato Coal Measures mudstone, sample 10290

TABLE VIII.2. XRF partial chemical analyses of kaolinite and montmorillonite material isolated from selected samples.

BULK CHEMICAL COMPOSITION OF ACID INSOLUBLE RESIDUES OF ONSHORE SAMPLES

The bulk chemical composition of the acid insoluble fraction of all samples is given in Tables VIII.3 and VIII.4.

APPENDIX VIII. BULK CHEMICAL COMPOSITION

ONSHORE SAMPLES

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10372	55	Rc	s	19.77	60.56	2.32	6.02	0.82	0.124	5.85	95.46	3.06
10371	61	Rc	s	14.04	56.04	2.20	8.89	1.05	0.163	4.72	87.10	3.99
10370	66	Rc	s	25.16	58.50	2.16	6.10	1.02	0.156	7.29	100.38	2.32
10369	78	Rc	s	18.90	62.90	2.47	0.82	0.96	0.047	6.27	92.36	3.33
10368	70	Rc	s	14.57	59.12	2.32	1.50	0.89	0.080	5.93	84.41	4.05
10367	21	Rc	s	21.73	56.49	2.08	0.15	1.04	0.022	6.88	88.39	2.60
10366	21	Rc	s	27.59	60.30	2.41	0.15	1.11	0.011	3.66	95.27	2.19
10365	8	Rc	s	13.98	44.50	0.56	0.01	1.85	0.044	12.22	73.16	3.18
10364	8	Rc	s	18.14	54.35	3.04	0.15	1.29	0.103	8.31	85.38	3.00
10363	86	Hw	sd	14.82	65.71	1.69	4.18	0.52	0.068	4.53	91.51	4.43
10362	86	Hw	sd	19.65	61.87	2.05	2.78	0.75	0.053	5.45	92.60	3.15
10361	85	Hw	sd	15.25	65.06	1.83	3.10	0.64	0.060	5.26	91.20	4.27
10360	81	Hw	sd	17.40	72.74	1.28	3.75	0.20	0.046	2.41	97.82	4.18
10359	86	Hw	md	18.58	59.40	1.11	5.24	0.64	0.073	6.31	91.35	3.20
10358	85	Hw	md	23.84	64.19	2.38	2.08	0.80	0.047	5.92	99.25	2.69
10357	81	Hw	md	18.90	61.39	3.00	1.75	0.80	0.067	6.23	92.13	3.25
10356	83	Wc	sst	19.44	68.01	2.45	1.16	0.73	0.049	4.56	96.39	3.50
10355	84	Wc	I.sst	22.12	62.28	3.01	1.12	0.89	0.047	5.57	95.03	2.82
10354	84	Wc	I.mst	21.16	59.82	3.45	0.94	1.00	0.050	6.34	92.76	2.83
10353	80	Wn	c.sst	19.22	62.69	3.31	1.55	0.72	0.063	5.64	93.19	3.26
10352	79	Wn	sst	24.81	68.88	2.88	1.22	0.70	0.045	5.15	103.68	2.78
10351	79	Wn	mst	20.94	67.44	2.82	1.14	0.74	0.050	5.11	98.24	3.22
10350	79	Wn	mst	18.47	67.02	2.88	1.15	0.84	0.047	5.19	95.59	3.63
10349	69	Tg	I.sst	20.62	74.40	2.77	1.22	0.59	0.042	3.81	103.45	3.61
10348	69	Tg	I.mst	19.98	61.70	3.27	1.11	0.87	0.059	6.31	93.29	3.09
10347	71	Tg	mst	22.02	60.27	2.98	1.16	0.83	0.062	5.67	92.99	2.74

TABLE VIII.3. Bulk chemical composition (weight %) of acid insoluble residues of sediments as determined by XRF analysis.

TABLE VIII.3. (continued).

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10346	75	Tg	mst	17.72	68.50	2.68	1.03	0.60	0.045	4.27	94.84	3.87
10345	77	Tg	mst	17.29	66.36	3.02	1.40	0.74	0.055	5.15	94.01	3.84
10344	72	Tg	mst	19.76	65.76	3.17	1.17	0.83	0.048	5.94	96.67	3.33
10343	76	Tg	mst	22.66	66.51	3.22	1.32	0.81	0.059	5.93	100.50	2.94
10342	74	Tg	mst	15.68	63.61	3.28	1.24	0.92	0.057	6.52	91.30	4.06
10341	68	Tg	mst	23.09	65.11	3.32	1.48	0.86	0.065	6.48	100.40	2.82
10340	70	Tg	mst	19.87	67.27	2.99	1.31	0.84	0.057	5.94	98.27	3.39
10339	73	Tg	con	12.78	56.03	2.47	6.96	0.51	0.050	4.17	83.42	4.38
10338	75	Tg	con	19.44	73.49	2.40	1.04	0.51	0.046	3.68	100.60	3.78
10337	72	Tg	con	22.55	62.82	3.21	1.43	0.84	0.061	6.05	96.96	2.79
10336	70	Tg	con	13.85	55.57	2.51	8.53	0.59	0.054	4.47	85.57	4.01
10335	67	Mt	sst	17.72	68.67	2.63	1.41	0.71	0.059	4.99	96.18	3.88
10334	65	Mt	sst	20.02	69.00	2.63	1.40	0.78	0.057	4.88	98.76	3.45
10333	63	Mt	sst	18.90	65.18	2.82	1.46	0.84	0.059	5.75	95.00	3.45
10332	64	Mt	sst	19.87	64.10	3.34	1.59	0.87	0.071	6.49	96.83	3.23
10331	62	Mt	sst	22.34	67.41	2.96	1.57	0.81	0.067	5.58	100.73	3.02
10330	67	Mt	con	20.84	63.69	2.84	2.82	0.77	0.063	5.55	96.57	3.06
10329	65	Mt	con	16.97	63.86	2.66	3.54	0.75	0.067	5.45	93.30	3.76
10328	64	Mt	con	17.40	68.22	2.64	1.68	0.72	0.061	5.17	95.89	3.92
10327	50	Ur	sst	16.32	58.12	0.44	3.13	0.73	0.030	6.38	85.15	3.56
10326	60	Ur	mst	21.48	67.54	3.05	1.48	0.89	0.066	5.74	100.24	3.14
10325	59	Ur	mst	22.12	64.41	3.43	1.50	0.90	0.072	6.37	98.80	2.91
10324	50	Ur	mst	20.08	68.00	2.93	1.59	0.84	0.066	5.75	99.25	3.39
10323	46	Ur	mst	21.48	65.41	3.61	1.36	0.96	0.078	7.01	99.90	3.05
10322	59	Ur	con	18.04	64.62	2.96	2.24	0.80	0.071	5.87	94.60	3.58
10321	50	Ur	con	16.43	69.30	2.91	1.75	0.85	0.071	5.87	97.18	4.22
10320	46	Ur	con	13.63	66.84	3.04	1.72	0.90	0.075	6.26	92.46	4.90

TABLE VIII.3 (continued).

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10319	59	Mt.M	sst	15.36	73.96	2.24	1.28	0.78	0.059	4.83	98.50	4.82
10318	46	Mt.M	sst	16.43	76.90	2.34	1.53	0.67	0.061	4.77	102.70	4.68
10317	42	Mt.M	sst	21.06	73.96	2.56	1.57	0.75	0.065	5.27	105.22	3.51
10316	57	Tw	mst	22.98	60.75	3.18	1.38	0.97	0.068	6.62	95.94	2.64
10315	51	Tw	mst	19.44	67.57	2.88	1.13	0.08	0.058	5.46	97.33	3.48
10314	57	Tw	con	18.04	63.07	3.05	2.01	0.90	0.064	6.17	93.30	3.50
10313	51	Tw	con	13.64	63.45	2.64	5.20	0.67	0.059	4.88	90.53	4.65
10312	39	Fy	sst	19.87	70.30	2.79	1.04	0.83	0.058	4.93	99.81	3.54
10311	38	Fy	I.sst	10.77	66.34	1.85	1.66	0.69	0.065	4.53	85.90	6.16
10310	41	Fy	I.sst	19.12	67.96	2.86	1.08	0.73	0.050	4.94	96.74	3.55
10309	56	Fy	I.sst	20.19	69.14	2.84	1.40	0.78	0.067	5.81	100.22	3.42
10308	43	Fy	I.sst	17.61	69.96	2.85	0.75	0.73	0.045	4.98	96.92	3.97
10307	38	Fy	I.mst	12.28	57.35	2.50	1.65	0.99	0.080	6.40	81.25	4.67
10306	41	Fy	I.mst	26.40	67.75	2.89	1.04	0.74	0.055	5.11	103.98	16.70
10305	56	Fy	I.mst	20.08	64.27	3.43	1.47	1.04	0.077	7.12	97.48	3.20
10304	43	Fy	I.mst	23.31	65.24	3.24	0.83	0.95	0.050	5.92	99.54	2.80
10303	32	Om	sst	18.33	63.77	2.14	1.78	0.08	0.070	4.73	90.90	3.48
10302	32	Om	I.sst	9.63	50.50	0.68	5.51	0.60	0.095	5.39	72.40	5.24
10301	36	Om	I.sst	15.49	58.63	2.72	1.56	0.88	0.066	5.81	85.15	3.79
10300	32	Om	I.mst	12.28	54.35	2.52	2.18	0.85	0.066	6.21	78.45	4.43
10299	36	Om	I.mst	15.12	62.70	2.42	1.75	0.91	0.078	5.99	88.96	4.15
10298	36	Om	con	14.74	57.78	2.24	1.45	1.00	0.069	6.22	83.49	3.92
10297	31	Pu	I.sst	0.00	46.22	0.61	10.07	0.56	0.165	4.61	62.23	0.00
10296	31	Pu	I.mst	0.00	44.94	2.30	3.67	0.53	0.075	5.20	56.79	0 00
10295	38	Pu	I.sst	2.83	56.07	1.11	4.41	0.84	0.069	6.04	71.36	19.81
10294	38	Pu	I.mst	18.33	53.50	2.48	0.92	1.02	0.072	7.60	83.92	2.92

TABLE VIII.3. (continued).

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10293	38	Mg	c.sst	10.58	53.92	1.80	3.05	0.89	0.061	6.44	76.74	5.10
10292	35	Mg	c.sst	17.61	72.95	2.75	0.83	0.63	0.046	4.23	99.04	4.14
10291	48	U.Mo	cgl	14.61	78.14	2.43	0.61	0.59	0.043	4.31	100.73	5.35
10290	48	U.Mo	sst	23.63	70.34	3.35	0.77	0.86	0.043	5.54	100.53	2.98
10289	31	U.Mo	c.sst	0.00	32.10	2.50	0.96	0.69	0.036	7.23	43.51	0.00
10288	33	U.Mo	sst	12.66	57.78	2.29	1.17	0.90	0.060	6.13	80.99	4.56
10287	33	U.Mo	mst	12.66	60.35	2.20	1.32	0.81	0.059	5.32	82.71	4.77
10286	38	U.Mo	mst	14.17	53.92	1.78	3.05	0.89	0.057	6.46	80.32	3.81
10285	21	U.Mo	mst	17.38	68.05	2.57	0.75	0.74	0.043	3.85	93.38	3.92
10284	44	U.Mo	mst	18.47	70.23	3.01	0.90	0.81	0.049	5.23	98.69	3.80
10283	33	U.Mo	con	15.12	60.77	2.04	2.24	0.79	0.061	5.41	86.43	4.02
10281	47	L.Mo	mst	15.57	85.78	2.84	0.35	0.64	0.015	2.30	107.49	5.51
10280	33	L.Mo	sst	16.44	63.77	2.35	1.32	0.82	0.042	4.28	89.02	3.88
10279	33	L.Mc	sst	14.17	58.20	2.19	1.56	0.91	0.053	5.60	82.68	4.11
10278	22	L.Mo	sst	18.14	59.92	2.03	0.78	0.81	0.057	5.29	87.02	3.30
10277	16	L.Mo	sst	13.61	69.76	1.99	1.36	0.70	0.060	4.76	92.24	5.13
10276	21	L.Mo	sst	16.82	64.62	2.41	0.66	0.79	0.033	4.05	89.38	3.84
10275	54	L.Mo	sst	18.26	86.64	2.85	0.42	0.60	0.018	2.53	111.31	4.74
10274	22	L.Mo	con	16.63	59.92	2.41	1.95	0.92	0.049	6.11	87.98	3.60
10273	21	L.Mo	I.sst	15.31	65.06	2.00	1.10	1.04	0.054	6.12	90.68	4.25
10272	21	L.Mo	I.sst	19.84	59.49	2.59	0.63	0.99	0.031	7.86	91.43	3.00
10271	27	Tmr	I.sst	21.91	67.69	2.84	0.48	0.76	0.036	5.50	99.21	3.09
10270	45	Tmr	I.sst	20.84	71.47	2.79	0.50	0.64	0.041	4.73	101.01	3.43
10269	58	Tmr	I.sst	18.04	65.18	2.11	0.46	0.76	0.043	4.83	91.41	3.61
10268	27	Tmr	I.mst	22.45	62.33	3.35	0.51	1.06	0.047	7.10	96.84	2.78
10267	45	Tmr	I.mst	20.30	62.53	3.28	0.40	1.12	0.051	6.76	94.44	3.08
10266	58	Tmr	I.mst	23.84	56.49	2.60	0.50	1.13	0.050	6.64	91.25	2.37

TABLE VIII.3. (continued).

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10265	16	Tmt	c.mst	16.05	60.77	2.39	1.27	1.03	0.046	6.56	88.11	3.79
10264	19	Tmt	c.mst	13.98	58.20	2.48	1.15	1.03	0.047	8.13	85.01	4.16
10263	20	Tmt	c.mst	14.36	58.20	2.70	0.90	1.08	0.050	6.82	84.11	4.05
10262	21	Tmt	mst	17.19	63.34	2.61	0.65	1.03	0.036	4.46	89.31	3.68
10261	30	Tmt	c.mst	9.45	58.20	2.15	1.13	0.87	0.057	5.82	77.67	6.16
10260	26	Tmt	c.mst	15.49	58.20	2.38	1.02	0.97	0.063	6.05	84.17	3.76
10259	30	Tmt	con	14.36	57.35	2.02	3.09	0.77	0.056	5.41	83.05	3.99
10258	26	Tmt	con	9.45	51.78	2.14	3.89	0.84	0.060	6.03	74.19	5.48
10257	30	Tmt	c.sst	13.23	53.50	2.29	1.17	0.87	0.057	5.54	76.65	4.04
10256	30	Tmt	c.sst	11.72	59.92	2.29	1.11	0.80	0.058	5.41	81.30	5.11
10255	30	Tmt	lst	1.32	53.92	2.15	2.84	0.95	0.042	7.42	68.64	40.85
10254	30	Tmt	lst	12.66	57.35	2.14	0.72	0.87	0.045	6.62	80.40	4.53
10253	30	Tmt	lst	13.23	55.21	2.46	1.90	0.87	0.048	6.37	80.08	4.17
10252	34	Tmt	lst	15.12	53.92	2.61	1.60	0.99	0.047	9.74	83.75	3.57
10251	34	Tmt	sty	17.57	62.49	2.54	0.81	0.96	0.060	5.51	89.94	3.56
10250	15	Ot	lst	7.56	43.23	2.22	3.62	0.75	0.045	12.38	69.80	5.72
10249	15	Ot	lst									
10248	15	Ot	lst									
10247	28	Ot	lst	0.00	77.04	1.64	0.70	0.60	0.046	4.78	84.80	0.00
10246	24	Ot	lst	0.00	50.08	2.66	3.15	0.86	0.045	8.57	65.36	0.00
10245	24	Ot	lst	0.00	47.93	2.19	5.47	0.78	0.034	8.18	64.58	0.00
10244	15	Ot	c.sst	9.26	53.93	2.91	1.68	0.74	0.043	9.59	78.15	5.82
10242	15	Or	lst	10.39	56.49	3.09	1.67	0.45	0.021	4.66	76.77	5.44
10241	28	Or	lst	9.07	60.78	2.25	2.97	0.46	0.028	5.01	80.56	6.70
10240	28	Or	lst	14.36	72.33	3.23	1.27	0.25	0.010	2.15	93.60	5.04
10239	23	Mp	sst	17.08	74.41	2.40	0.48	0.67	0.030	5.33	100.40	4.36
10238	23	Mp	sst	15.04	86.92	1.92	0.23	0.57	0.020	3.45	108.15	5.78

TABLE VIII.3. (continued).

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10236	4	T.A	mst	0.00	57.57	1.02	0.65	0.55	0.022	4.43	64.24	0.00
10235	4	T.A	mst	8.50	71.04	1.70	0.91	0.72	0.026	4.35	87.23	8.36
10234	4	T.A	mst	0.00	51.36	0.98	0.52	0.57	0.024	3.83	56.83	0.00
10233	15	T.A	lst	0.00	50.50	2.17	0.50	0.09	0.021	3.01	56.29	0.00
10232	9	T.A	lst	3.21	56.92	1.86	4.88	0.73	0.021	4.64	72.26	17.37
10231	9	T.A	lst	13.61	58.20	2.25	1.92	1.00	0.023	3.97	80.97	4.28
10230	28	T.A	lst	19.28	73.19	3.95	1.29	0.33	0.018	2.58	100.63	3.80
10229	15	Ao	c.sst	0.00	20.12	1.14	0.57	0.53	0.033	7.07	29.46	0.00
10227	11	Ao	c.sst	0.00	52.43	1.88	1.38	0.82	0.040	5.96	62.51	0.00
10226	28	Ao	c.sst	16.82	78.32	3.39	0.75	0.16	0.016	2.43	101.79	4.66
10225	28	Ao	sst	13.60	81.32	2.59	0.29	0.20	0.039	4.29	102.32	5.98
10224	28	Wh	lst	11.34	54.78	1.95	2.16	0.92	0.016	7.35	78.51	4.83
10223	4	Wh	c.sst	10.02	71.69	1.06	1.04	0.64	0.022	3.67	88.14	7.15
10222	7	Wh	c.mst	8.50	66.34	1.50	1.33	0.81	0.024	4.51	83.01	7.80
10221	11	Wh	c.mst	8.88	66.76	1.95	1.66	0.83	0.033	5.21	85.32	7.52
10220	4	Wh	c.mst	0.00	32.74	0.74	0.37	0.50	0.021	4.11	38.48	0.00
10219	28	Wh	c.mst	5.86	65.27	1.43	1.75	0.90	0.029	4.24	79.47	11.14
10218	5	G.M	lst	4.16	45.58	2.77	5.08	0.44	0.033	14.26	72.32	10.96
10215	5	Mk	mst	16.06	54.35	1.63	0.58	1.06	0.036	6.23	79.94	3.38
10213	17	Mk	mst	17.95	53.92	1.25	0.10	1.24	0.015	4.39	78.86	3.00
10212	17	Mk	mst	0.00	16.69	0.55	0.34	0.87	0.033	5.27	23.75	0.00

TABLE VIII.3. (continued).

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10211	4	Wk	mst	20.40	52.43	0.47	0.05	2.03	0.000	3.13	78.51	2.57
10210	1	Wk	mst	20.97	57.78	0.66	0.22	1.60	0.001	1.53	82.76	2.76
10209	28	Wk	mst	21.92	47.51	0.69	0.04	2.47	0.000	2.41	75.04	2.17
10208	28	Wk	mst	22.49	54.35	1.66	0.26	1.96	0.000	1.98	82.70	2.42
10207	18	Wk	mst	23.24	54.35	1.61	0.16	1.50	0.011	2.10	82.97	2.34
10206	23	Wk	mst	21.48	73.62	1.58	0.11	1.11	0.010	1.58	99.49	3.43
10205	20	Wk	sst	14.82	97.46	1.05	0.09	0.57	0.020	1.58	115.59	6.58
10204	13	Js	mst	13.20	61.20	2.35	2.26	0.88	0.082	5.94	85.91	4.64
10203	12	Js	mst	9.71	29.96	2.53	0.46	0.55	0.048	4.69	47.94	3.09
10202	12	Js	con	9.54	24.18	1.01	3.31	1.00	0.041	3.30	42.38	2.53
10201	28	Ts	sst	16.44	56.49	2.34	1.51	0.96	0.096	6.49	84.32	3.44

RIVER AND HARBOUR SAMPLES.

Sample	Local.	Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10396	87	Rc		13.49	66.08	2.72	2.02	0.80	0.071	6.39	91.57	4.89
10395	82	Rc		13.72	71.51	2.26	1.65	0.67	0.062	4.72	94.59	5.21
10394	57	Rc		7.88	71.90	2.42	1.23	0.55	0.049	3.85	87.87	9.12
10393	58	Rc		20.20	62.33	2.62	1.64	0.85	0.068	6.31	94.01	3.08
10392	58	Rc		18.58	66.74	2.48	1.72	0.78	0.070	5.48	95.85	3.59
10391	55	Rc		15.55	53.71	1.80	6.72	1.59	0.175	10.83	90.37	3.45
10390	61	Rc		13.82	60.49	2.36	6.52	0.90	0.141	6.33	90.56	4.37
10389	66	Rc		15.67	59.98	2.26	7.17	0.92	0.148	6.49	92.63	3.82
10388	70	Rc		13.38	67.88	2.14	2.65	0.95	0.093	6.01	93.10	5.07
10387	52	Rc		22.13	59.89	2.52	1.67	0.84	0.083	6.18	93.31	2.70

TABLE VIII.3. (continued).

Sample	Local.	Strat.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
10386	53	Rc	18.47	64.77	2.52	1.58	0.73	0.066	5.32	93.45	3.50
10385	49	Rc	18.35	63.02	2.56	1.46	0.82	0.075	5.93	92.21	3.43
10384	40	Rc	12.40	63.36	2.12	3.04	0.87	0.106	6.75	88.64	5.10
10383	37	Rc	16.84	61.09	2.12	1.86	1.04	0.095	7.69	90.73	3.62
10382	35	Rc	12.97	50.95	1.74	3.75	1.77	0.172	13.08	84.43	3.92
10381	29	Rc	16.63	60.69	2.24	1.15	0.85	0.074	6.22	87.85	3.64
10380	6	Rc	16.52	63.30	3.14	1.81	0.78	0.074	10.11	95.82	3.83
10379	6	Rc	17.18	59.17	2.24	1.15	1.10	0.049	7.98	89.26	3.44
10378	6	Rc	14.36	69.48	1.92	1.82	1.24	0.065	5.83	94.71	4.83
10377	6	Rc	16.42	55.78	1.84	0.55	1.19	0.041	7.46	83.28	3.39
10376	12	Rc	20.09	62.14	2.50	1.30	0.94	0.074	8.12	95.16	3.09
10375	12	Rc	18.26	64.11	1.94	4.58	0.90	0.121	6.66	96.57	3.51
10374	10	Rc	13.17	68.00	1.56	1.71	1.15	0.112	8.98	94.68	5.16
10373	10	Rc	14.36	61.97	1.96	5.75	0.84	0.129	6.23	91.23	4.31

WESTERN SHELF SAMPLES

Sample	Strat.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
A45	Rc	14.56	61.14	2.50	2.23	0.87	0.070	6.21	87.58	4.31
A47	Rc	13.29	64.05	2.69	2.10	0.94	0.065	6.59	89.72	4.81
A51	Rc	19.16	61.27	3.28	1.89	0.95	0.061	6.81	93.42	3.19
B319	Rc	13.40	72.03	2.50	1.81	0.75	0.061	5.45	96.00	5.37
B321	Rc	21.60	59.02	3.61	1.47	0.90	0.063	6.72	93.38	2.73
B322	Rc	18.90	70.62	3.05	1.57	0.77	0.063	5.25	100.22	3.73
B645	Rc	21.49	55.25	3.60	1.43	0.94	0.063	8.22	90.99	2.57
B647	Rc	14.04	70.14	2.92	1.47	0.73	0.067	4.86	94.22	4.99
B648	Rc	13.93	70.79	2.62	2.16	0.55	0.063	5.59	95.70	5.08
B649	Rc	17.71	52.85	2.64	3.39	0.89	0.085	10.32	87.88	2.98

TABLE VIII.3. (continued).

Sample	Strat.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
B651	Rc	7.88	68.69	1.44	4.48	0.60	0.098	6.62	89.80	8.71
B653	Rc	5.84	31.15	1.14	3.21	0.68	0.378	22.21	64.60	5.33
B667	Rc	7.35	43.91	1.04	3.01	5.88	0.266	19.89	81.34	5.97
B675	Rc	15.33	67.38	2.58	1.62	0.90	0.055	5.92	93.78	4.39
B676	Rc	17.41	67.17	1.84	3.33	1.64	0.132	7.94	99.46	3.85
B680	Rc	23.11	62.29	2.98	2.21	0.87	0.061	6.38	97.90	2.69
B681	Rc	17.60	67.83	2.86	1.82	0.77	0.062	5.52	96.46	3.85
B682	Rc	20.2	64.49	3.00	1.72	0.84	0.065	6.01	96.37	3.19
B778	Rc	13.61	64.60	2.14	1.93	0.80	0.065	6.84	89.98	4.74
B780	Rc	9.24	63.77	2.20	1.88	0.78	0.057	6.23	84.15	6.90
B781	Rc	12.85	69.74	2.16	2.41	0.78	0.071	5.59	93.60	5.42
B784	Rc	12.23	71.77	2.22	2.48	1.15	0.103	5.59	95.54	5.86
B788	Rc	10.57	67.85	1.92	3.32	0.78	0.083	6.09	90.61	6.41
B780	Rc	9.24	63.77	2.20	1.88	0.78	0.057	6.23	84.15	6.90
B781	Rc	12.85	69.74	2.16	2.41	0.78	0.071	5.59	93.60	5.42
B784	Rc	12.23	71.77	2.22	2.48	1.15	0.103	5.59	95.54	5.86
B788	Rc	10.57	67.85	1.92	3.32	0.75	0.083	6.09	90.61	6.41
B793	Rc	12.63	73.08	2.50	1.57	0.70	0.054	4.49	95.02	5.79
B795	Rc	15.99	79.47	2.36	1.57	0.62	0.059	4.18	104.24	4.97
B796	Rc	16.20	66.63	2.40	2.14	0.77	0.066	6.05	94.25	4.11
B799	Rc	14.04	63.21	1.88	4.72	1.12	0.129	8.31	93.40	4.50
B801	Rc	10.91	62.06	2.66	2.65	0.84	0.074	6.54	85.73	5.69
B802	Rc	16.52	58.42	2.62	2.79	0.85	0.068	8.61	89.87	3.54
B804	Rc	17.60	62.31	2.22	3.92	0.82	0.086	6.61	93.56	3.54
B812	Rc	18.14	58.93	3.00	1.93	0.94	0.063	6.95	89.95	3.25
B813	Rc	10.79	58.95	2.80	2.32	0.92	0.066	6.91	82.75	5.46
B816	Rc	16.20	61.88	3.12	1.71	0.87	0.059	6.72	90.55	3.82
B818	Rc	15.55	68.22	2.66	1.97	0.63	0.057	5.15	94.23	4.39

TABLE VIII.3. (continued).

Sample	Strat.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
B822	Rc	15.12	71.60	2.94	1.43	0.72	0.061	4.83	96.70	4.74
C169	Rc	23.76	59.06	3.42	1.53	0.94	0.059	6.88	95.64	2.49
C170	Rc	15.88	70.40	2.80	1.55	0.80	0.066	4.99	96.48	4.43
C171	Rc	21.81	60.06	3.28	1.75	0.97	0.075	6.61	94.55	2.75
C172	Rc	18.03	56.34	3.38	1.78	0.90	0.063	6.85	87.34	3.12
C173	Rc	21.58	60.17	3.24	1.65	0.85	0.059	6.44	93.98	2.79
C174	Rc	18.14	64.52	3.18	1.67	0.85	0.058	6.19	94.60	3.56
C175	Rc	19.67	59.74	2.64	3.21	0.92	0.081	6.66	92.92	3.04
C178	Rc	11.66	61.18	1.06	8.34	1.00	0.179	8.54	91.95	5.25
C262	Rc	13.40	70.64	1.94	2.94	0.60	0.074	4.39	93.98	5.27
C265	Rc	15.01	68.00	2.04	2.58	0.67	0.071	5.23	93.60	4.53
C272	Rc	14.46	73.80	2.10	2.11	0.84	0.102	4.85	98.26	5.10
C275	Rc	13.82	75.67	2.06	1.81	0.65	0.059	4.18	98.24	5.48
C276	Rc	13.15	79.50	2.18	1.39	0.52	0.052	3.69	100.48	6.05
C277	Rc	17.14	73.05	2.36	1.48	0.68	0.052	4.35	99.11	4.26
C281	Rc	10.24	68.33	2.08	2.90	0.60	0.066	4.66	95.54	6.67
C293	Rc	14.48	71.60	2.00	2.79	0.63	0.077	5.15	96.72	4.94
C330	Rc	10.57	65.91	2.16	2.21	0.95	0.095	5.75	87.64	6.23
C332	Rc	11.55	69.03	1.80	4.51	1.00	0.125	5.59	93.60	5.97
C340	Rc	13.06	66.01	1.96	2.46	1.54	0.148	6.19	91.36	5.05
C348	Rc	11.89	71.71	2.04	3.49	0.62	0.084	4.49	94.32	6.03
C355	Rc	16.31	73.03	1.94	2.80	1.09	0.125	5.65	100.94	4.47
C363	Rc	13.49	83.20	2.08	1.62	0.72	0.072	3.75	104.93	6.16
C364	Rc	16.95	80.46	2.10	1.48	0.60	0.055	3.53	90.17	4.74
C368	Rc	14.46	78.34	2.22	1.64	0.77	0.074	4.06	101.56	5.41
C372	Rc	14.48	72.73	2.34	2.04	0.85	0.080	4.85	97.37	5.02

APPENDIX VIII BULK CHEMICAL COMPOSITION

ONSHORE SAMPLES

Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
Hw	sd	16.78	66.34	1.71	3.45	0.53	0.057	4.41	93.28	3.93
Hw	md	20.44	61.66	2.16	3.02	0.75	0.062	6.15	94.24	3.02
Wc	sst	20.78	65.14	2.73	1.14	0.81	0.048	5.06	95.71	3.13
Wc	mst	21.16	59.82	3.45	0.94	1.00	0.050	6.34	92.76	2.83
Wn	sst	22.01	65.78	3.09	1.38	0.71	0.054	5.39	98.41	2.99
Wn	mst	19.70	67.23	2.85	1.15	0.79	0.048	5.15	96.92	3.41
Tg	I.sst	20.62	74.40	2.77	1.22	0.59	0.042	3.81	103.45	3.61
Tg	I.mst	19.98	61.70	3.27	1.11	0.87	0.059	6.31	93.30	3.09
Tg	mst	19.76	65.42	3.08	1.26	0.80	0.056	5.74	96.12	3.31
Mt	sst	19.77	66.87	2.88	1.49	0.80	0.063	5.54	97.41	3.38
Ur	sst	16.32	58.12	0.14	3.13	0.73	0.030	6.38	84.85	3.56
Ur	mst	21.29	66.34	3.25	1.48	0.90	0.071	6.22	99.55	3.12
Mt.M	sst	17.62	74.94	2.38	1.46	0.73	0.062	4.96	102.15	4.25
Tw	mst	11.49	64.16	3.03	1.25	0.88	0.063	6.04	86.91	5.58

TABLE VIII.4. Average bulk chemical composition (weight %) of acid insoluble residues determined by XRF analysis for major lithotypes in the onshore study areas.

TABLE VIII.4. (continued).

Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
Fy	sst	19.87	70.30	2.79	1.04	0.83	0.058	4.93	99.82	3.54
Fy	I.sst	16.19	68.48	2.51	1.27	0.73	0.059	5.12	94.36	4.23
Fy	I.mst	18.55	62.29	3.06	1.32	0.99	0.069	6.48	92.98	3.36
Om	sst	18.33	63.77	2.14	1.78	0.08	0.070	4.73	90.90	3.48
Om	I.sst	12.56	54.56	1.70	3.53	0.74	0.081	5.60	78.77	4.34
Om	I.mst	13.70	58.52	2.47	1.96	0.88	0.072	6.10	83.70	4.27
Pu	I.sst	1.41	51.14	0.86	7.24	0.70	0.117	5.32	66.79	36.27
Pu	I.mst	9.16	49.22	2.39	2.29	0.77	0.074	6.44	70.34	5.37
Mg	c.sst	14.09	63.43	2.27	1.94	0.76	0.053	5.33	87.87	4.50
U.Mo	cgl	14.61	78.14	2.43	0.61	0.59	0.043	4.31	100.73	5.35
U.Mo	sst	12.10	53.41	2.71	2.90	0.82	0.046	6.30	78.28	4.41
U.Mo	mst	15.67	63.14	2.39	4.47	0.81	0.052	5.21	91.74	4.03
L.Mo	mst	15.57	85.78	2.84	0.35	0.64	0.015	2.30	107.49	5.51
L.Mo	sst	16.24	67.15	2.30	1.02	0.77	0.044	4.42	91.94	4.13
L.Mo	I.sst	15.31	65.06	2.00	1.10	1.04	0.054	6.12	90.68	4.25
L.Mo	I.mst	19.84	59.49	2.59	0.63	0.99	0.031	7.86	91.43	3.00
Tmr	I.sst	19.98	68.07	2.65	0.63	0.72	0.043	5.00	97.09	3.41
Tmr	I.mst	23.25	62.27	3.03	0.61	1.01	0.051	6.40	96.62	2.68
Tmt	c.mst	14.53	59.61	2.42	1.03	0.98	0.055	6.67	85.29	4.67
Tmt	c.sst	12.47	56.71	2.29	1.13	0.83	0.057	5.47	78.95	4.55
Tmt	lst	10.58	55.10	2.34	1.76	0.92	0.045	7.53	78.27	5.21
Ot	lst	1.89	54.57	2.18	3.23	0.75	0.043	8.48	71.14	28.87
Ot	c.sst	9.26	53.93	2.91	1.68	0.74	0.043	9.59	78.15	5.82

TABLE VIII.4. (continued).

Strat.	Lith.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{SiO_2}{Al_2O_3}$
Or	lst	11.27	63.2.	2.86	1.97	0.39	0.020	3.94	83.65	5.61
Mp	sst	16.06	80.66	2.16	0.35	0.62	0.025	4.39	104.26	5.02
T.A	mst	2.83	59.99	1.23	0.69	0.61	0.024	4.20	69.57	21.20
Ao	lst	9.02	59.70	2.56	2.15	0.54	0.021	3.55	77.54	6.62
Ao	c.sst	5.60	50.29	2.14	0.90	0.50	0.030	5.15	64.61	8.98
Ao	sst	13.60	81.32	2.59	0.29	0.20	0.039	4.29	102.33	5.98
Wh	lst	11.34	54.78	1.95	2.16	0.92	0.016	7.35	78.52	4.83
Wh	c.sst	10.02	71.69	1.06	1.04	0.64	0.022	3.67	88.14	6.32
Wh	c.mst	5.81	57.78	1.41	1.28	0.76	0.027	4.52	71.32	9.94
G.M	lst	4.16	45.58	2.77	5.08	0.44	0.033	14.26	72.32	10.96
Mk	mst	17.01	54.14	1.44	0.34	1.00	0.026	5.31	79.27	3.18
Wk	mst	21.75	56.67	1.11	0.14	1.78	0.004	2.12	83.57	2.61
Wk	sst	14.82	97.46	1.05	0.09	0.57	0.020	1.58	115.59	6.58
Js	mst	11.46	45.58	2.44	1.36	0.72	0.060	5.32	66.94	3.98
Ts	sst	16.44	56.49	2.34	1.51	0.96	0.096	6.49	84.33	3.44

TABLE VIII.4. (continued).

RIVER AND HARBOUR SAMPLES

Local.	Strat.	Al_2O_3	SiO_2	K_2O	CaO	TiO_2	MnO	Fe_2O_3	Total	$\frac{Al_2O_3}{SiO_2}$
6	Rc	16.12	61.93	2.28	1.33	1.08	0.057	7.84	90.64	3.87
10	Rc	13.76	64.98	1.76	3.73	0.99	0.120	7.60	92.94	4.73
12	Rc	19.17	63.12	2.22	2.94	0.92	0.097	7.39	95.87	3.30
29	Rc	16.63	60.69	2.24	1.15	0.85	0.074	6.22	87.85	3.64
35	Rc	12.97	50.95	1.74	3.75	1.77	0.172	13.08	84.43	3.92
37	Rc	16.84	61.09	2.12	1.86	1.04	0.095	7.69	90.73	3.62
40	Rc	12.40	63.36	2.12	3.04	0.87	0.106	6.75	88.64	5.10
49	Rc	18.35	63.02	2.56	1.46	0.82	0.075	5.93	92.21	3.43
50	Rc	22.13	59.89	2.52	1.67	0.84	0.083	6.18	93.31	2.70
53	Rc	18.47	64.77	2.52	1.58	0.73	0.066	5.32	93.45	3.50
55	Rc	15.55	53.71	1.80	6.72	1.59	0.175	10.83	90.37	3.45
57	Rc	7.88	71.90	2.42	1.23	0.55	0.049	3.85	87.87	9.12
58	Rc	19.39	64.53	2.55	1.68	0.81	0.069	5.89	94.93	3.34
61	Rc	13.82	60.49	2.36	6.52	0.90	0.141	6.33	90.56	4.37
66	Rc	15.67	59.98	2.26	7.17	0.92	0.148	6.49	92.63	3.82
70	Rc	13.38	67.88	2.14	2.65	0.95	0.093	6.01	93.10	5.07
82	Rc	13.72	71.51	2.26	1.65	0.67	0.062	4.72	94.59	5.21
87	Rc	13.49	66.08	2.72	2.02	0.80	0.071	6.39	91.57	4.89

Variations in the chemical composition (Figs. VIII.2A and VIII.2B can be related to mineralogy.

High SiO_2/Al_2O_3 ratios in the Oligocene sediments result from a high quartz to feldspar ratio of samples together with the predominance of Si-rich montmorillonite clay minerals (Table VIII.2) in the clay fraction. The high SiO_2/Al_2O_3 ratios in volcanogenic sandstones reflect the significant increase in plagioclase feldspar relative to clay minerals in these samples. Moreover, the occurrence of α-cristobalite in some Oligocene and most volcanogenic sediments would further enhance the SiO_2/Al_2O_3 ratio. Coal measure mudstones and other kaolinite-rich samples exhibit a low SiO_2/Al_2O_3 ratio because of their high kaolinite content.

Compared to the Oligocene and volcanogenic sediments, post-Mahoenui Group and Mesozoic sediments contain relatively more K_2O as a result of their higher illite content. In glauconitic Oligocene sediments however, the K_2O values are higher.

The typically high CaO content in the insoluble residues from limestones compared to mudstones and sandstones probably reflects increased Ca^{2+} absorption in the clay minerals of limestones. High CaO values in volcanogenic sandstones, however, reflect the abundant calcic-plagioclase in these samples.

High TiO_2 values in coal measure sediments are a function of the stability of Ti under conditions of acid lessivage.

The Fe_2O_3 content of Oligocene sediments is variable and, as with K_2O, may be a function of the glauconite content

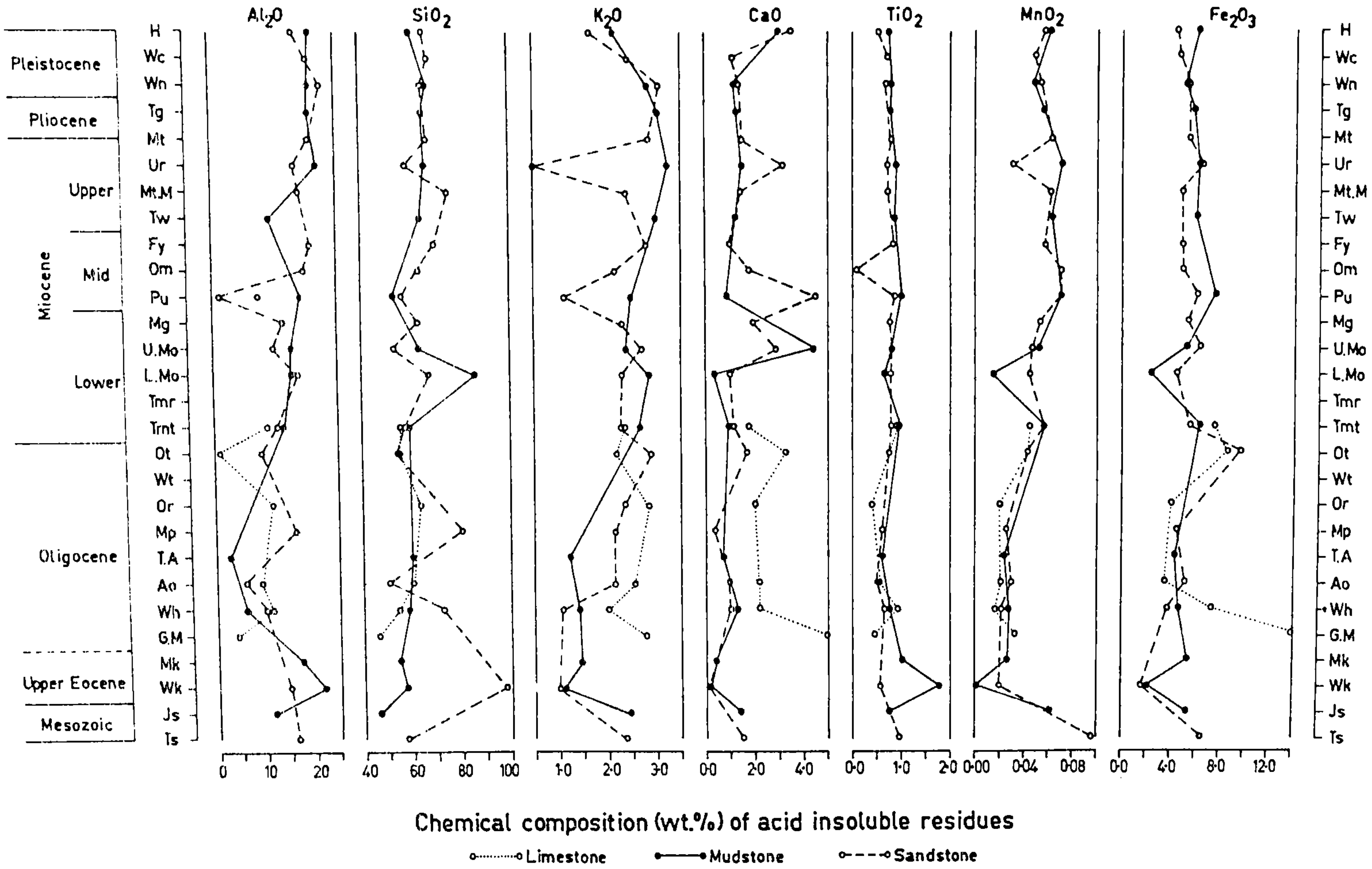


Fig. VIII.2A. Bulk chemical composition of acid insoluble residues of the major limstone, mudstone and sandstone units in the onshore study area as determined by XRF analysis. Abbreviations of formation names defined in Table 1.1.

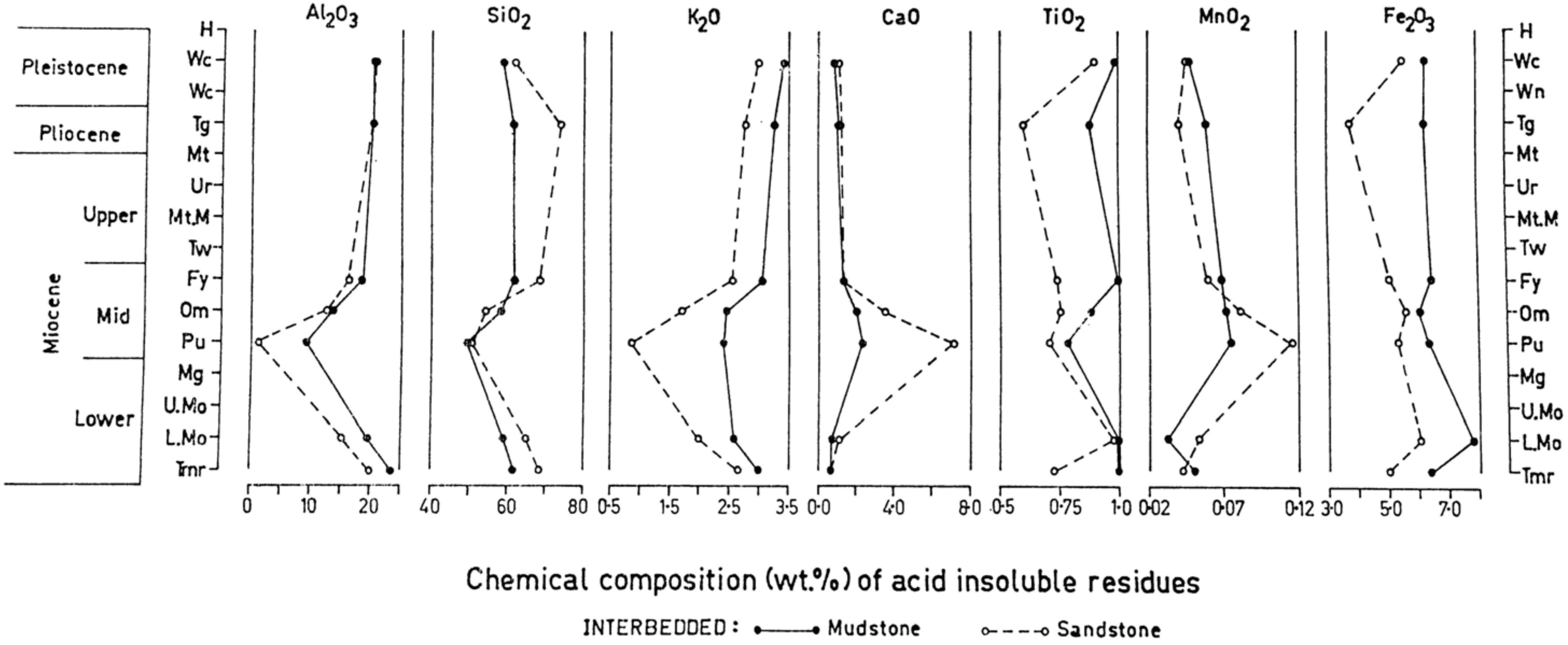


Fig. VIII.2B. Bulk chemical composition of acid insoluble residues of the major interbedded mudstone and sandstone units in the onshore study area as determined by XRF analysis. Abbreviations of formation names defined in Table 1.1.

of these sediments. In Mesozoic and post-Miocene sediments the Fe_2O_3 content is fairly constant. In interbedded sandstone/mudstone sequences the greater quantity of clay minerals in the mudstone units compared to the sandstone units produces a lower SiO_2/Al_2O_3 ratio, a higher K_2O value (due to increase in illite), a higher Fe_2O_3 value (due to increase in chlorite), and higher TiO_2 and MnO_2 values (due to less leaching in the mudstones).

BULK CHEMICAL COMPOSITION OF ACID INSOLUBLE RESIDUES OF WESTERN SHELF SURFICIAL SEDIMENT SAMPLES

The bulk chemical composition of the acid insoluble fraction of all samples is given in Tables VIII.3 and VIII.4. Variations in the chemical composition of the acid insoluble residues on the shelf (Figs. VIII.3 to VIII.7) can be related to mineralogic variations.

High SiO_2/Al_2O_3 ratios (Fig. VIII.3) in sediments south of Cape Egmont, in the outer-Central Taranaki Shelf, and the outer-Hamilton Shelf, result from the large quartz/feldspar plus clay minerals ratio in these sediments. The low SiO_2/Al_2O_3 ratio in South Taranaki Shelf and North Cook Strait Basin sediments is a function of increasing clay mineral abundance.

K_2O is most abundant in mid- and outer-shelf positions and in sediments to the south (Fig. VIII.4). This is partly due to an increase in the content of illite in the clay size fraction in this region. Optical examination of the sand- and silt-size fraction of the sediments indicates the presence of appreciable quantities of micas (mostly sericite) that were probably derived from the erosion of the granite

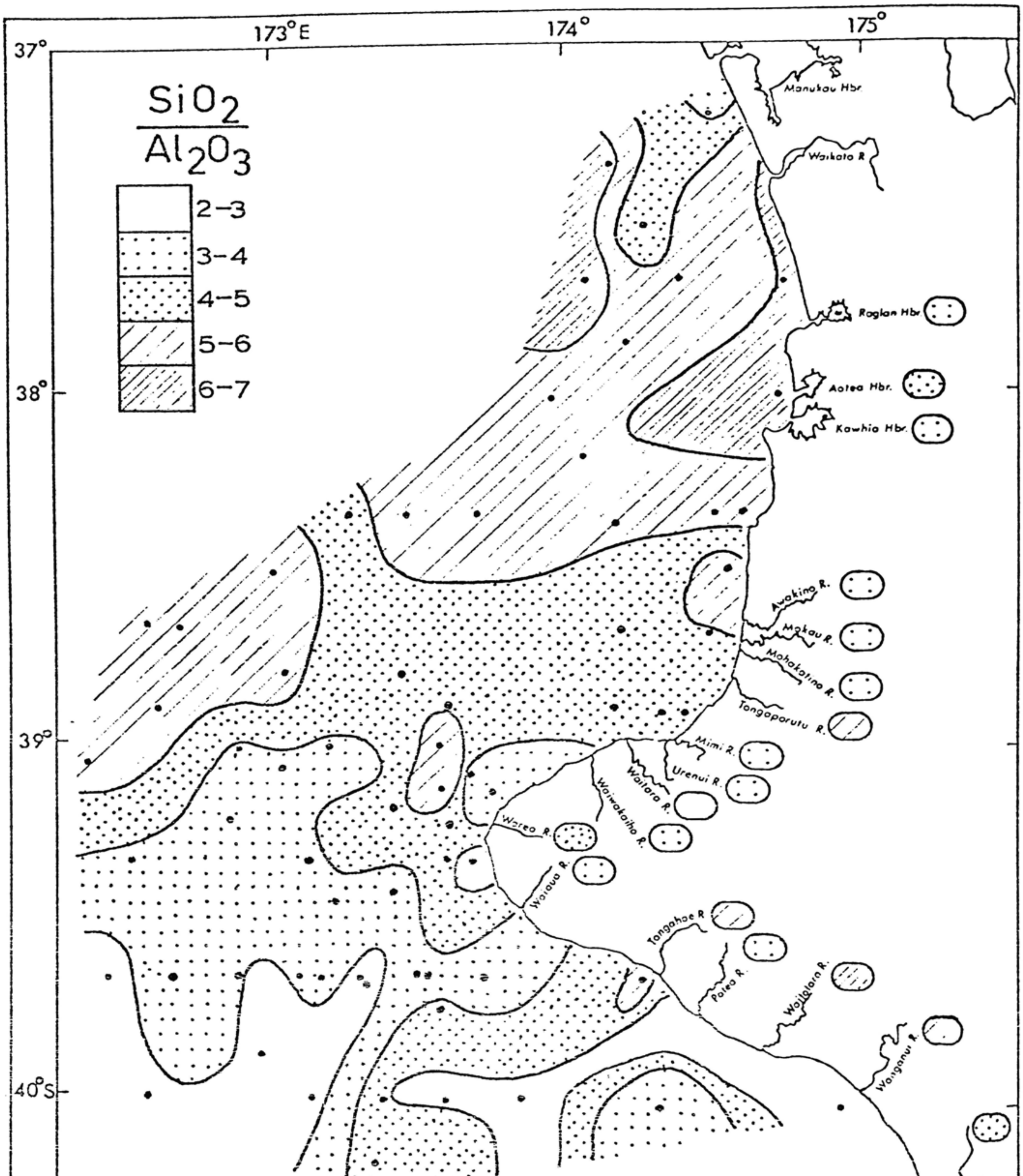


Fig. VIII.3. $\frac{SiO_2}{Al_2O_3}$ ratio in the acid insoluble residue from surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

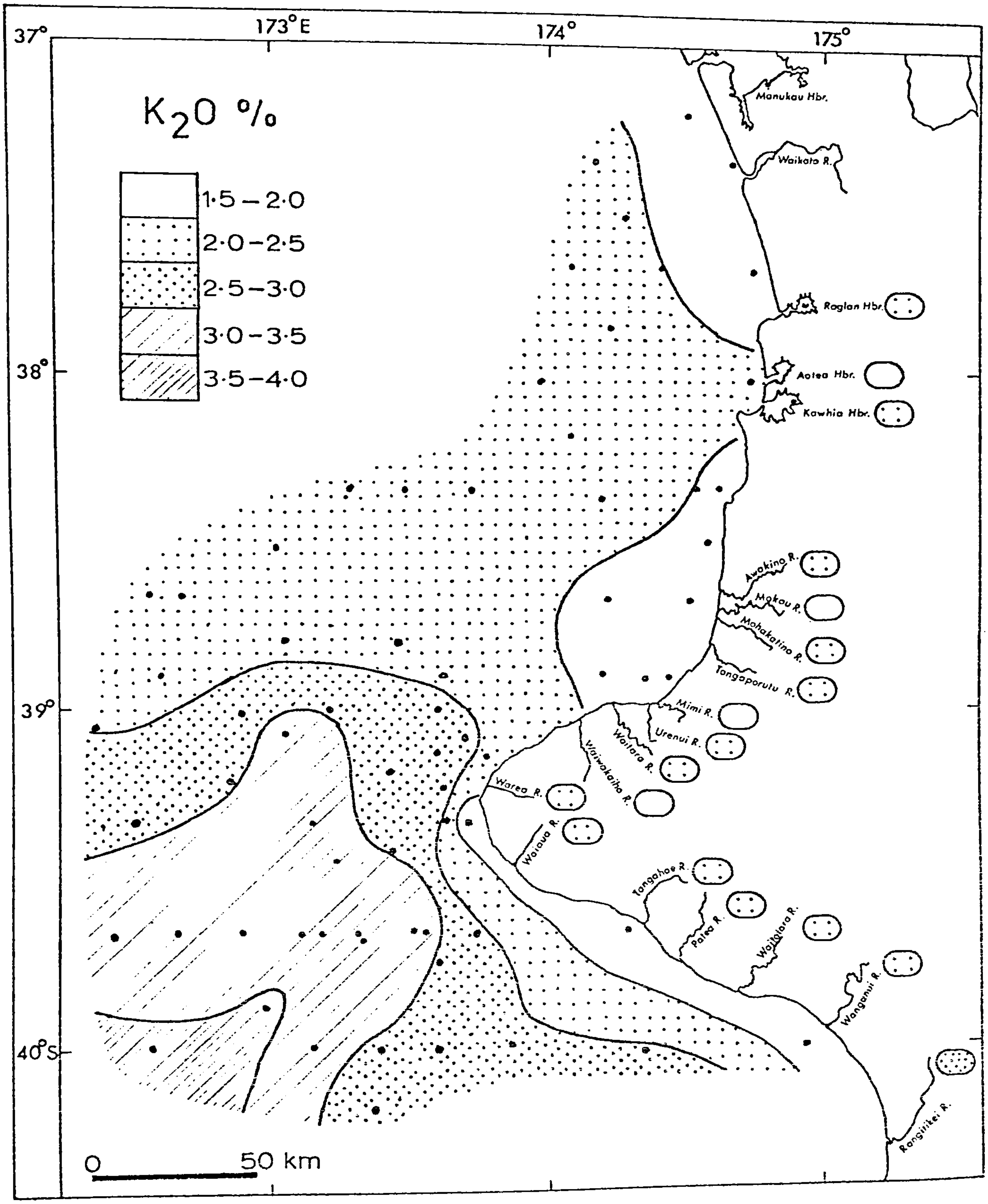


Fig. VIII.4. K_2O abundance (weight % in acid insoluble residue) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

rocks in the northwestern South Island and were subsequently carried north by the Westland Current.

The higher concentrations of CaO in nearshore sediments (Fig. VIII.5) reflects the abundance of calcic plagioclase derived from the erosion of Quaternary volcanic ash units that mantle much of the onshore area.

TiO_2 is expectedly most abundant in samples containing appreciable amounts of titanomagnetite. Vanadium shows a similar distribution.

High MnO values in nearshore sediments (Fig. VIII.6) parallel the high MnO content of soil materials developed on volcanogenic sediments in the neighbouring onshore areas.

Fe_2O_3 shows a more complex distribution pattern (Fig. VIII.7). High concentrations occur in some samples that contain appreciable amounts of titanomagnetite. The high content of Fe_2O_3 in South Taranaki Shelf sediments reflects the large proportions of chlorite in these sediments.

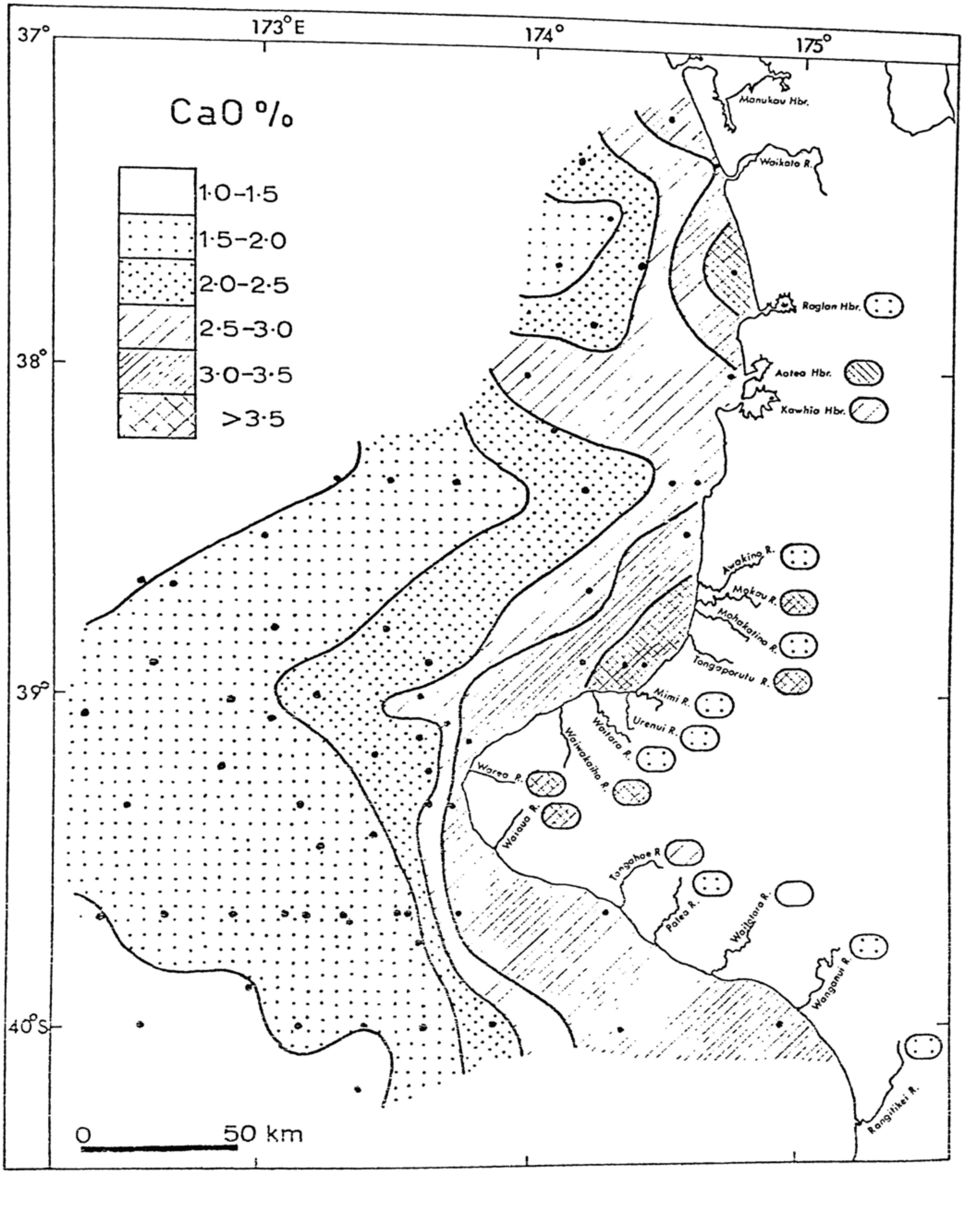


Fig. VIII.5. CaO abundance (weight % in acid insoluble residue) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

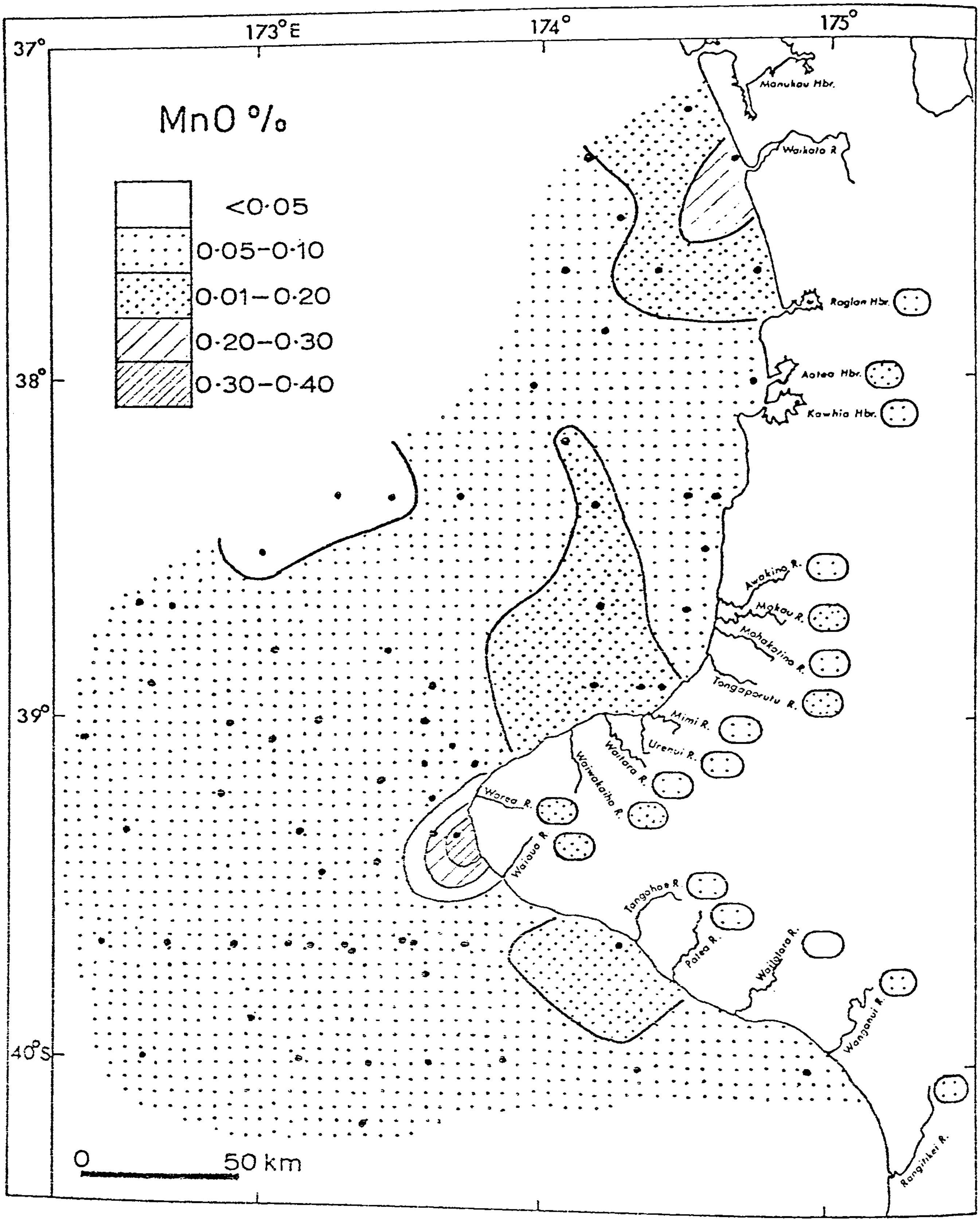


Fig. VIII.6. MnO abundance (weight % in acid insoluble residue) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

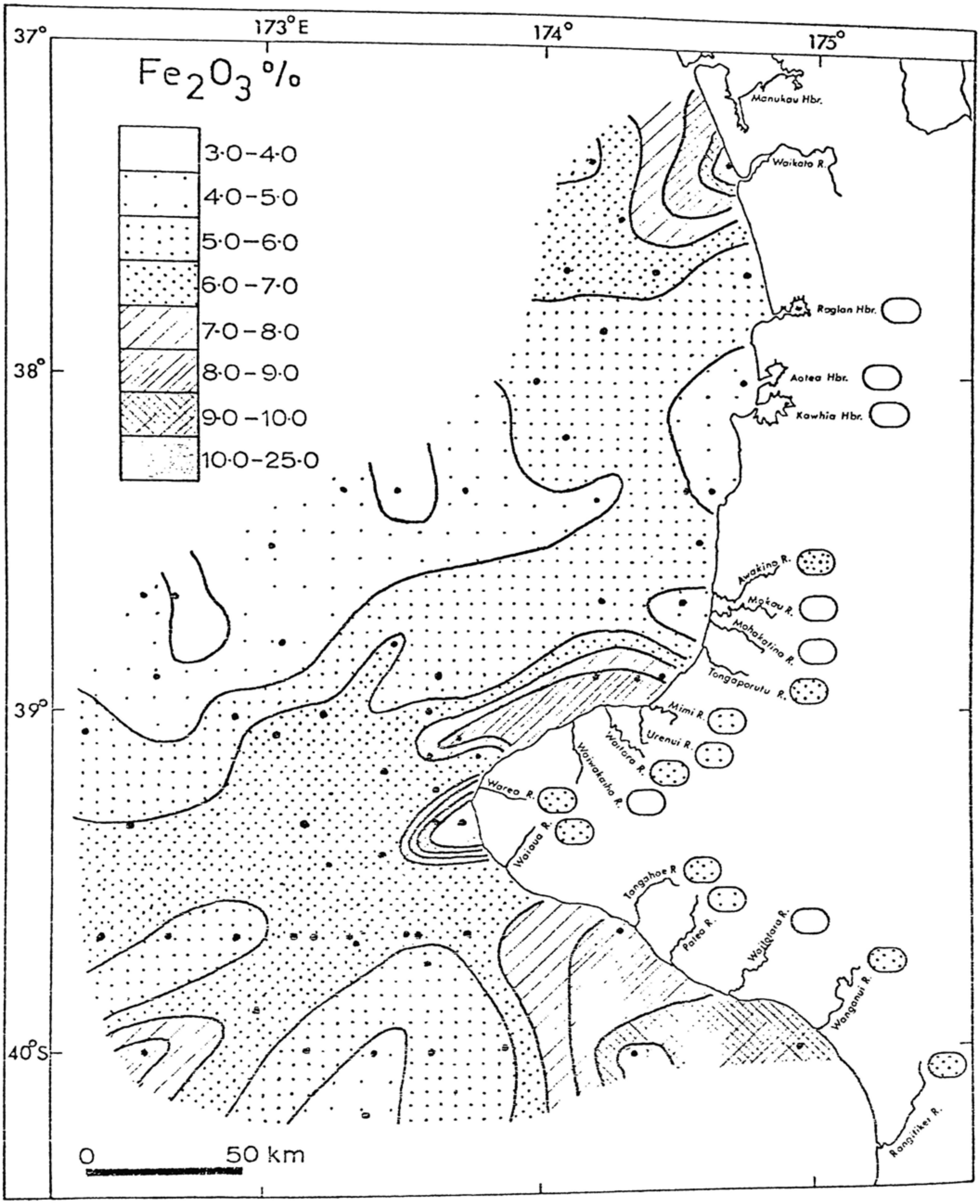


Fig. VIII.7. Fe_2O_3 abundance (weight % in acid insoluble residue) in surficial sediments of the western shelf and adjacent harbours and rivers. Harbour and river sediment data represent average values.

APPENDIX IX

SEDIMENTARY UNIT OR FORMATION	SYMBOL	N.Z.STAGE	THICKNESS (m)		BULK LITHOLOGY (%)			SEDIMENTATION PERIOD (m.y.)	SEDN.RATE(cm/1000 yr)	
			Max.	Av.	Mst.	Sst.	Lst.		Max.	Av.
Pleistocene	Pl	Wn-H	2260	1510	50	50	10	1.8	125.5	83.7
Tangahoe Mudstone Fm.	Tg	Wo-Wm	1340	875	95	5	-	2.5	53.6	35.0
Matemateaonga Sandstone Fm.	Mt	Tk-Wo	2750	1500	10	90	-	3.5	78.6	42.9
Urenui Siltstone Fm.	Ur	Up.Tt	500	400	90	10	-	1.5	33.3	26.7
Mt Messenger Sandstone Fm.	Mt.M	Low-Mid.Tt	800	600	25	75	-	2.5	32.0	24.0
Omoao Formation	Om	Sw-Low.Tt	150	100	50	50	-	2.5	6.0	4.0
Purupuru Volcanic Sandstone Fm.	Pu	Sc-Sw	70	50	38	60	2	3.5	2.0	1.4
Mangarara Sandstone Fm.	Mg	Mid-Up.Pl	63	20	5	90	5	5	1.3	0.4
Upper Mokau Sandstone Fm.	U.Mo	Mid.Pl	170	100	60	40	-	3	5.7	3.3
Lower Mokau Sandstone Fm.	L.Mo	Low-Mid.Pl	120	60	5	95	-	3	4.0	2.0
Taumarunui Fm.	Tmr	Po-Pl	1500	600	60	40	-	3.5	42.9	17.1
Taumatamaire Fm.	Tmt	Up.Lw-Po	300	150	90	5	5	3	10.0	5.0
Otorohanga Limestone Fm.	Ot	Lw	65	25	-	-	100	4.5	1.4	0.6
Orahiri Limestone Fm.	Or	Ld	60	25	-	-	100	3.5	1.7	0.7
Aotea Sandstone Fm.	Ao	Mid Lwh-Ld	120	45	15	75	10	5	2.4	0.9
Whaingaroa Siltstone Fm.	Wh	Low Mid.Lwh	180	30	80	10	10	4	4.5	0.8
Waikato Coal Measures Fm.	Wk	Low.Lwh	36	3	75	25	-	1	3.6	0.3

TABLE IX.1. Bulk variations in thickness, lithology and sedimentation rates for the major Cenozoic sedimentary units in the onshore study area.

APPENDIX IX

SEDIMENTARY UNIT OR FORMATION	SYMBOL	NO. OF SAMPLES ANALYSED	BULK MINERALOGY (wt. %)					RATIO	INSOLUBLE RESIDUE MINERALOGY (wt. %)			
			Qtz.	Plag.	Pot.	Clay	Calc.	qtz. / feld.	Qtz.	Plag.	Pot.	Clay
Pleistocene	Pl	14	22	30	3	35	3	0.7	23	31	3	36
Tangahoe Mudstone Fm.	Tg	14	26	27	3	42	2	0.8	26	28	3	43
Matemateaonga Sandstone Fm.	Mt	8	28	31	4	35	2	0.8	28	32	4	36
Urenui Siltstone Fm.	Ur	8	23	28	1	43	2	0.8	23	29	1	44
Mt Messenger Sandstone Fm.	Mt.M	3	37	33	4	25	1	1.0	35	32	4	26
Omoao Fm.	Om	16	25	30	2	35	7	0.8	27	33	2	37
Purupuru Volcanic Sandstone Fm.	Pu	4	12	42	2	32	6	0.3	13	45	2	34
Mangarara Sandstone Fm.	Mg	2	20	18	4	35	23	0.9	25	22	5	40
Upper Mokau Sandstone Fm.	U.Mo	7	30	25	2	36	7	1.1	32	27	2	38
Lower Mokau Sandstone Fm.	L.Mo	9	37	25	5	33	-	1.2	37	25	5	33
Taumarunui Fm.	Tmr	6	23	19	1	50	7	1.2	25	21	1	53
Taumatamairo Fm.	Tmt	19	14	10	1	43	32	1.2	21	15	1	63
Otorohanga Limestone Fm.	Ot	8	3	1	-	6	90	3.0	26	10	6	46
Orahiri Limestone Fm.	Or	9	8	6	2	6	78	1.0	34	46	21	24
Aotea Sandstone Fm.	Ao	36	20	13	5	12	50	1.1	39	18	23	32
Whaingaroa Siltstone Fm.	Wh	37	10	4	1	31	54	2.0	30	15	14	47
Waikato Coal Measures Fm.	Wk	11	39	3	1	57	-	9.8	39	3	1	57

TABLE IX.2. Bulk variations in mineral composition (average mineralogy weighted according to bulk lithology of formation) for the major Cenozoic sedimentary units in the onshore study area. Mineral abundance determined by XRD analysis.

SEDIMENTARY UNIT OR FORMATION	SYMBOL	NO. OF SAMPLES ANALYSED	CLAY MINERALOGY (wt.%)					CLAY MINERAL CRYSTALLINITY		
			Kaol.	Ill.	Mont.	Chlor.	M/L	Ill.	Chlor.	Mont.
Pleistocene Sediments	Pl	14	2	44	44	4	6	2.0	3.3	2.8
Tangahoe Mudstone	Tg	10	2	44	34	12	8	0.4	0.8	3.7
Matemateaonga Sandstone	Mt	5	1	44	29	22	4	0.4	0.4	3.3
Urenui Siltstone	Ur	5	1	47	30	18	4	0.3	0.4	4.3
Mt. Messenger Sandstone	Mt.M	3	2	57	12	16	13	0.4	0.8	1.1
Omoao Formation	Om	16	3	47	26	18	6	1.1	0.8	4.6
Purupuru Volcanic Sandstone	Pu	4	-	25	61	6	8	4.6	7.1	6.5
Mangarara Sandstone	Mg	2	1	25	58	-	16	2.0	1.9	7.1
Upper Mokau Sandstone	U.Mo	7	6	53	19	11	11	0.9	1.3	3.6
Lower Mokau Sandstone	L.Mo	9	10	43	24	9	14	0.9	2.2	3.6
Taumarunui Formation	Tmr	6	10	41	41	2	6	1.4	2.0	5.4
Taumatamaire Formation	Tmt	8	3	29	53	8	7	1.7	3.1	7.7
Otorohanga Limestone	Ot	7	2	22	70	2	4	3.7	8.5	5.2
Orahiri Limestone	Or	3	4	12	82	-	2	5.1	6.1	7.0
Aotea Sandstone	Ao	9	4	15	79	1	1	4.8	7.0	6.2
Whaingaroa Siltstone	Wh	5	1	7	91	1	-	7.5	9.8	7.4
Waikato Coal Measures	Wk	7	79	10	-	6	5	8.1	-	-

Table IX.3. Bulk variations in clay mineral composition (determined by XRD analysis of the <2μ size fraction) for the major Cenozoic sedimentary units in the onshore study area. Clay mineral abundance and crystallinity are weighted according to the bulk lithology of the sedimentary unit.

PHOTOGRAPHIC

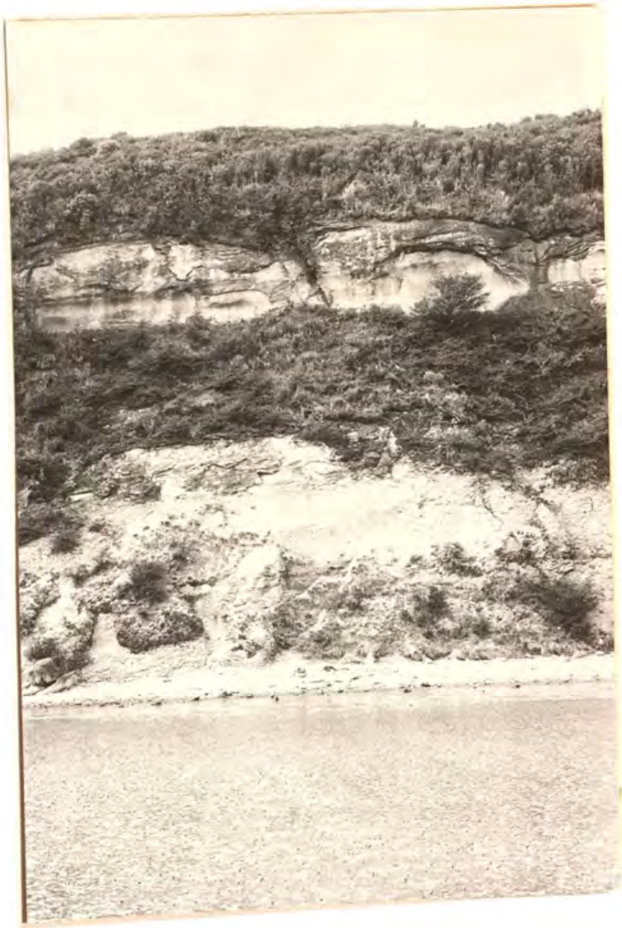
PLATES

PLATES

Plate 1.1. Coastal exposure of well-bedded, blocky, indurated Triassic (Oretian Stage) sandstone about 2 km south of the mouth of the Kiritehere Stream, Marakopa (N82/227801). The white bed at the base of the exposure is a tuffaceous sandstone.

Plate 1.2. Sediments of the Waikato Coal Measures exposed in an open-cast mine at Huntly (N56/694767; Locality 1). Pale kaolinitic mudstone is interbedded with dark bituminous coal seams containing fine kaolinitic laminations.

Plate 1.3. Type Section of the Aotea Sandstone on the south-east shore of Aotea Harbour (N73/379199; Locality 11). Lower frittered, blue-grey Whaingaroa Siltstone with occasional blocky fine sandstone beds passes gradationally upwards into more massive, iron-stained Aotea Sandstone. Scale provided by the pack on the beach.



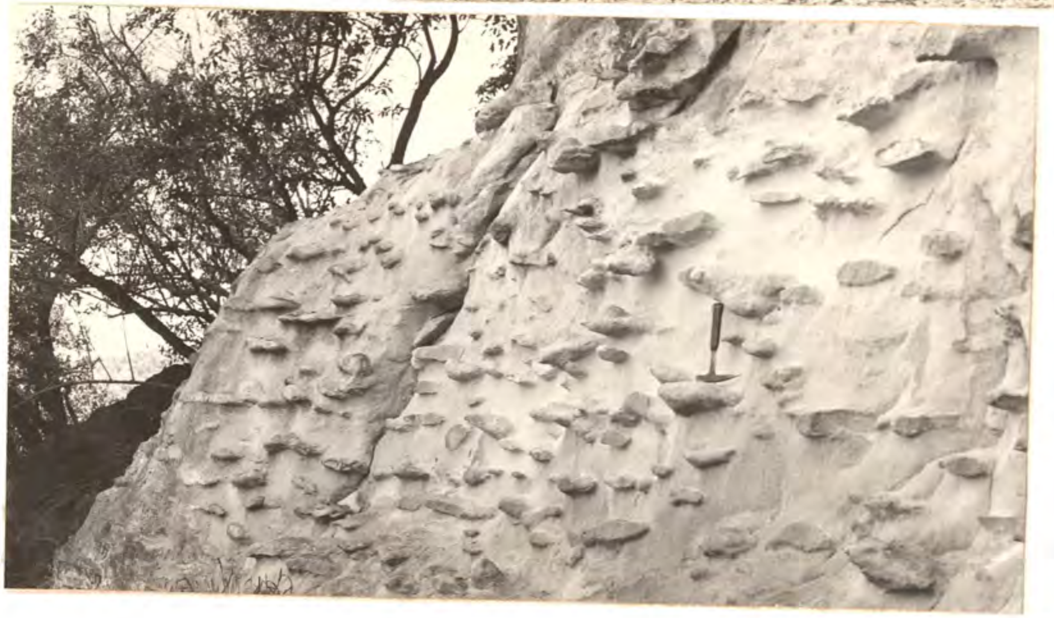


Plate 1.4. Massive, light blue-grey calcareous mudstones of the Whaingaroa Siltstone showing conchoidal fracture and frittered surface at the type locality on the eastern shore of Raglan Harbour (N64/492463).

Plate 1.5. The Banded Sandstone lithofacies of the Aotea Sandstone at N82/489773, northwest Mairoa, showing typical elliptical "concretions" of hard calcareous sandstone in a soft friable sand.

Plate 1.6. The Flaggy Limestone lithofacies of the Orahiri Limestone at N86/394722, near Waitanguru, showing a monotonous sequence of moderately to well developed flags averaging 8 cm thick.

Plate 1.7. Type section of the Orahiri Limestone (Or), Waitomo Sandstone (Wt) and Otorohanga Limestone (Ot) at the former Otorohanga Limestone Company Quarry, southwest of Otorohanga (N74/642943; Locality 15). The lower grass covered slope conceals Aotea Sandstone (Ao) which is in turn sharply overlain by flaggy Orahiri Limestone, massive Waitomo Sandstone and flaggy Otorohanga Limestone.

Plate 1.8. Cliff of Otorohanga Limestone in the north Piopio area (N83/541686), showing incipiently flagged, well flagged and more knobbly weathering characteristics in the limestone. Mudstones of the Taumatamaire Formation overlie the highest limestone exposures. Scale given by person sitting on ledge near centre.

Plate 1.9. Exposure of a more sandy phase of the Taumatamaire mudstone in the Awakino area (N91/303472; Locality 26) showing prominent bands of calcareous concretions up to 60 cm thick.


Ot
Wt
Or
Ao

Plate 1.10. Flysch-type sequence of the Taumarunui Formation exposed on the main highway immediately north of National Park (N111/885874; Locality 58). Sandstone bands form the more prominent but less abundant units. Note the small fault traversing the outcrop which is approximately 7 m high.

Plate 1.11. Typical bluff outcrop of massive, ferruginous Lower Mokau Sandstone, north of Awakino township. Photograph taken looking northeast from N91/181409.

Plate 1.12. Roadside exposure of massive, ferruginous Lower Mokau Sandstone, north of Awakino township (N91/185408).

Plate 1.13. Muddy sandstones with occasional calcareous concretions in the Lower Mokau Sandstone exposed 11 km north of Ohura (N91/496313).

Plate 1.14. Massive sandy mudstones of the Upper Mokau Sandstone are unconformably overlain by a conglomerate which is the local equivalent of the Mangarara Sandstone, and by an alternating sequence of sandstone and mudstone bands belonging to the Lower Purupuru Volcanic Sandstone. The conglomerate contains large clasts of Upper Mokau Sandstone. Exposure in Awakino area (N91/165402; Locality 31).

Plate 1.15. An impressive exposure of interbedded sandstone and mudstone of the Lower Purupuru Volcanic Sandstone at the mouth of the Mokau River (N91/160350). Person on beach provides scale.

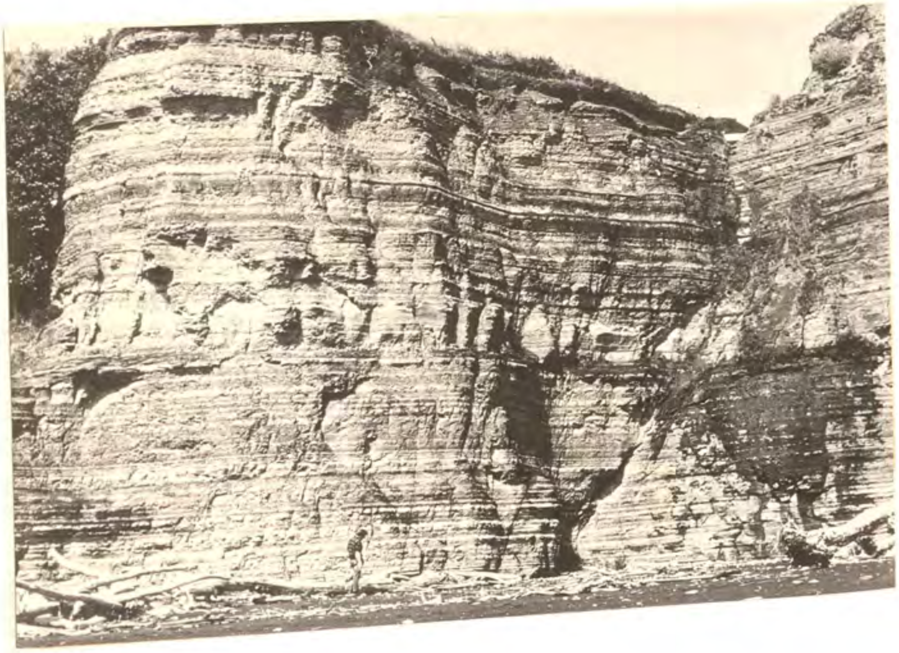


Plate 1.16. An exposure of Mohakatino beds at (N100/419193) in the Waitaanga Saddle area in which blue-grey mudstone is overlain by about 60 cm of green-grey tuffaceous sandstone (forming a prominent ledge) which in turn passes up into a massive muddy fine sandstone. The tuffaceous layer probably represents a tongue of the Purupuru Volcanic Sandstone. Compare the homogeneity of this inland exposure to the well bedded nature of the coastal outcrops in Pls. 1.14 and 1.15.

Plate 1.17. Exposure of Ferry Sandstone at the type locality on Mohakatino Road (N91/184300; Locality 38). Dark green-grey, massive, micaceous sandstone is interbedded with light blue-grey non-calcareous mudstone; both lithologies are non-tuffaceous.

Plate 1.18. Massive blue-grey mudstone of the uppermost Mohakatino Group (Tawariki Mudstone?) at N100/428193 in the Waitaanga Saddle area.

Plate 1.19. Typical bluff outcrop of Mt. Messenger Sandstone near Ahititi. Photograph taken looking east from N100/150170; Locality 42.

Plate 1.20. Grey mudstones and overlying massive, brown sandstones of the Mt. Messenger Sandstone, immediately north of the Tongaporutu River Mouth (N100/135223). Note scour contact between sandstone and mudstone lithologies.

Plate 1.21. Roadside exposure of Urenui mudstone, north of Urenui (N99/032041; Locality 50), showing concretionary horizons and a thin tuffaceous sandstone band about 1 m above the author's head.

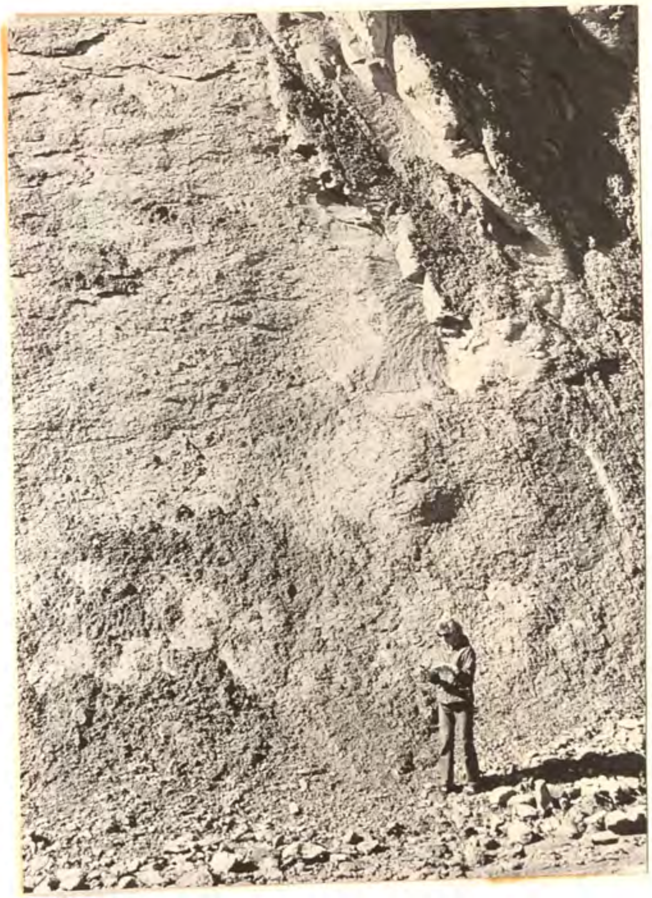


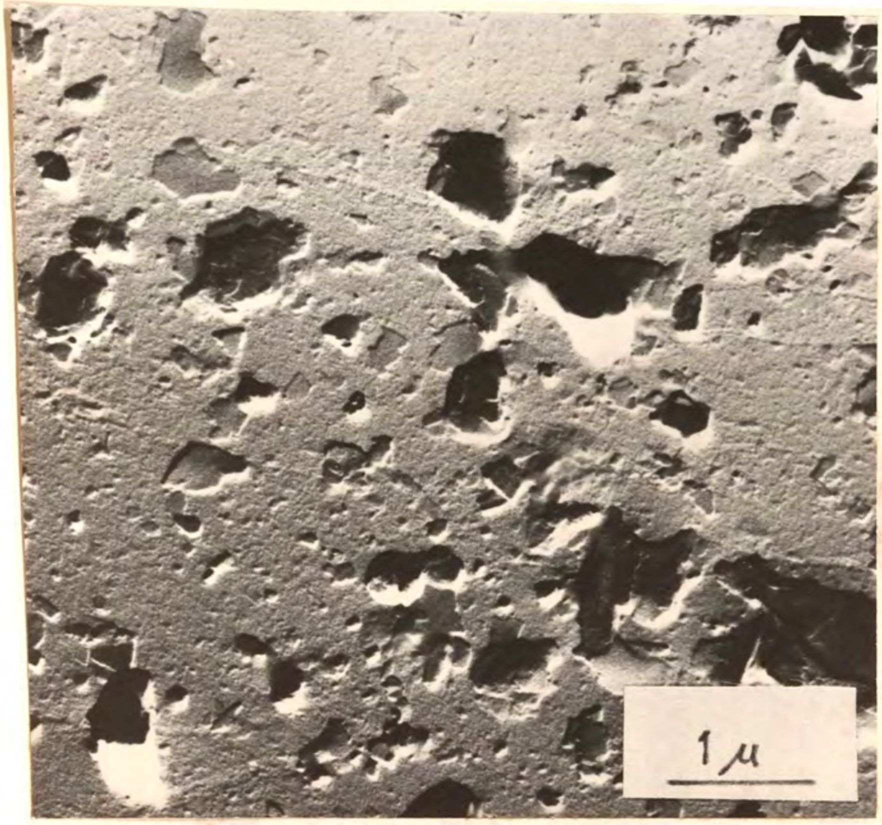
Plate 1.22. Exposure of massive Matemateaonga Sandstone with scattered calcareous concretions in the Strathmore Saddle area (N119/057637; Locality 64).

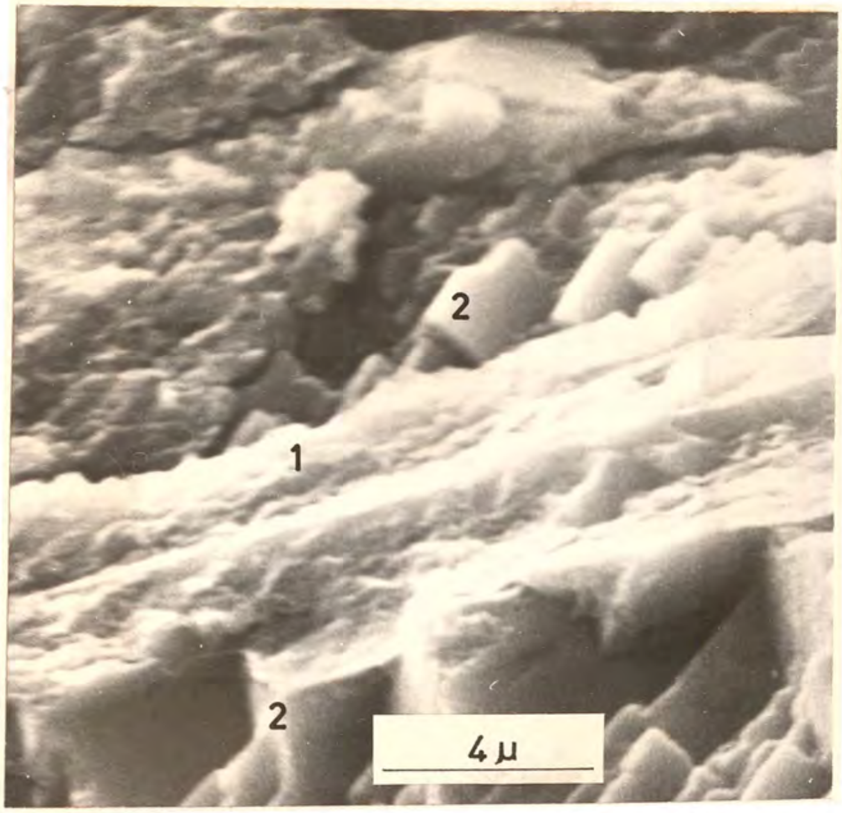
Plate 1.23. Massive, frittered, blue-grey Tangahoe Mudstone in a roadside outcrop on the main highway 7 km north of Taihape (N132/244239; Locality 71).

Plate 1.24. A sandy phase of the Tangahoe Mudstone exposed on the main highway 1 km south of Kakatahi (N131/838206; Locality 72). Compare the less frittered appearance of this outcrop to that in Plate 1.23 and note the horizon of large calcareous concretions 4 m above car.

Plate 3.1. Transmission electron micrograph of the clay fraction (<2µ) of a Triassic sandstone (10201). Note the angular anhedral to subhedral form of the illitic, chloritic and mixed-layer clay minerals.

Plate 3.2. Scanning electron micrograph of the fractured surface of a calcareous concretion (10202), from the Jurassic beds at locality 12, showing layers of clay-rich material (montmorillonite) and large calcite crystals (2). The montmorillonitic seam material is considered to represent diagenetically altered volcanogenic material.


1 µ


2
1
2
4 µ

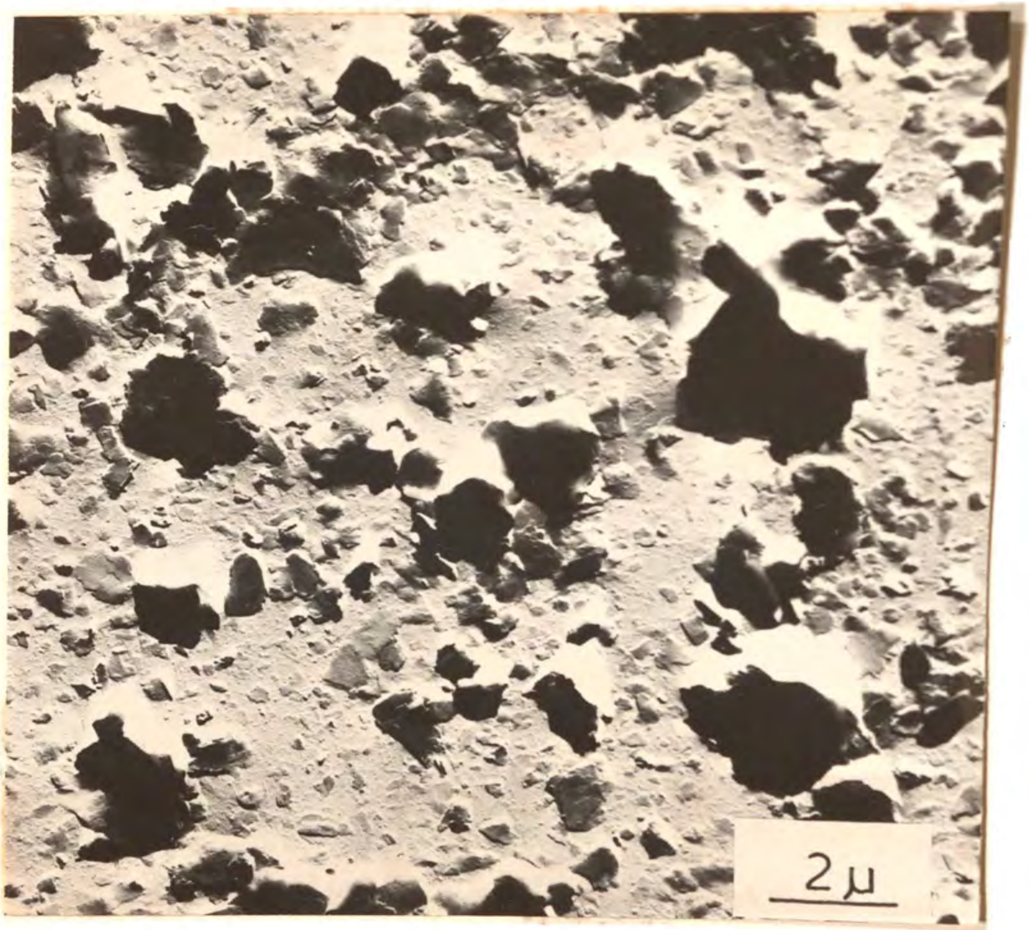
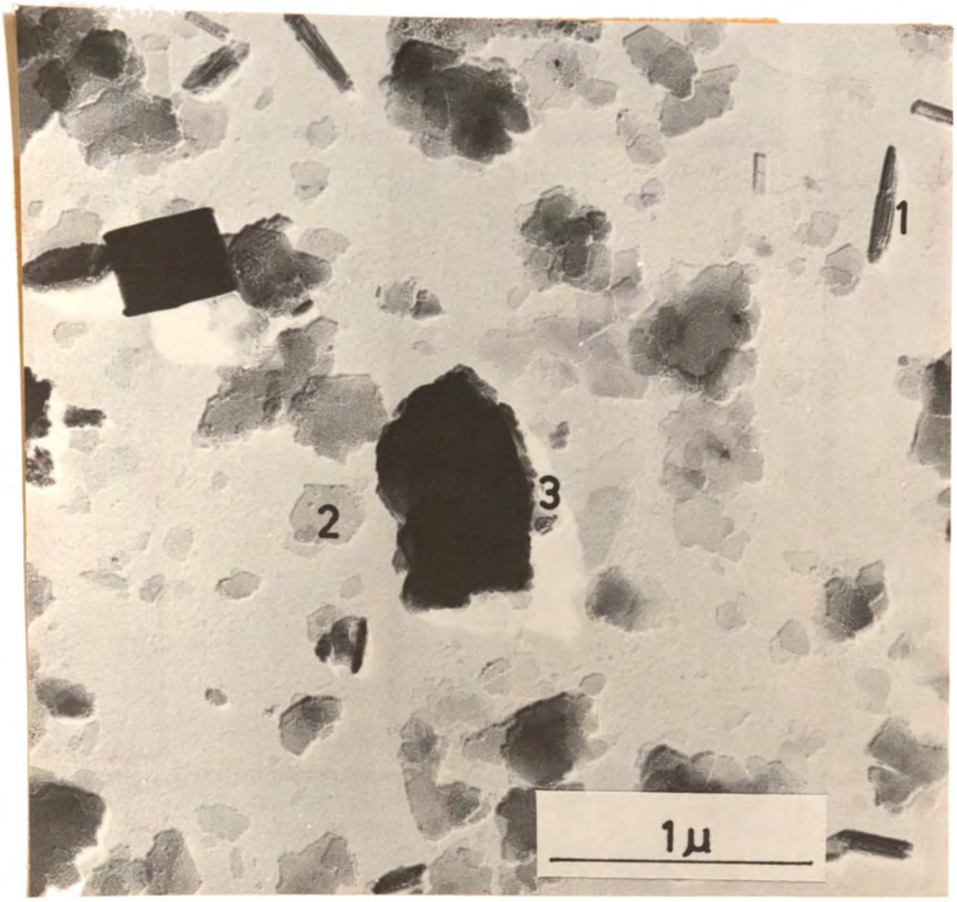
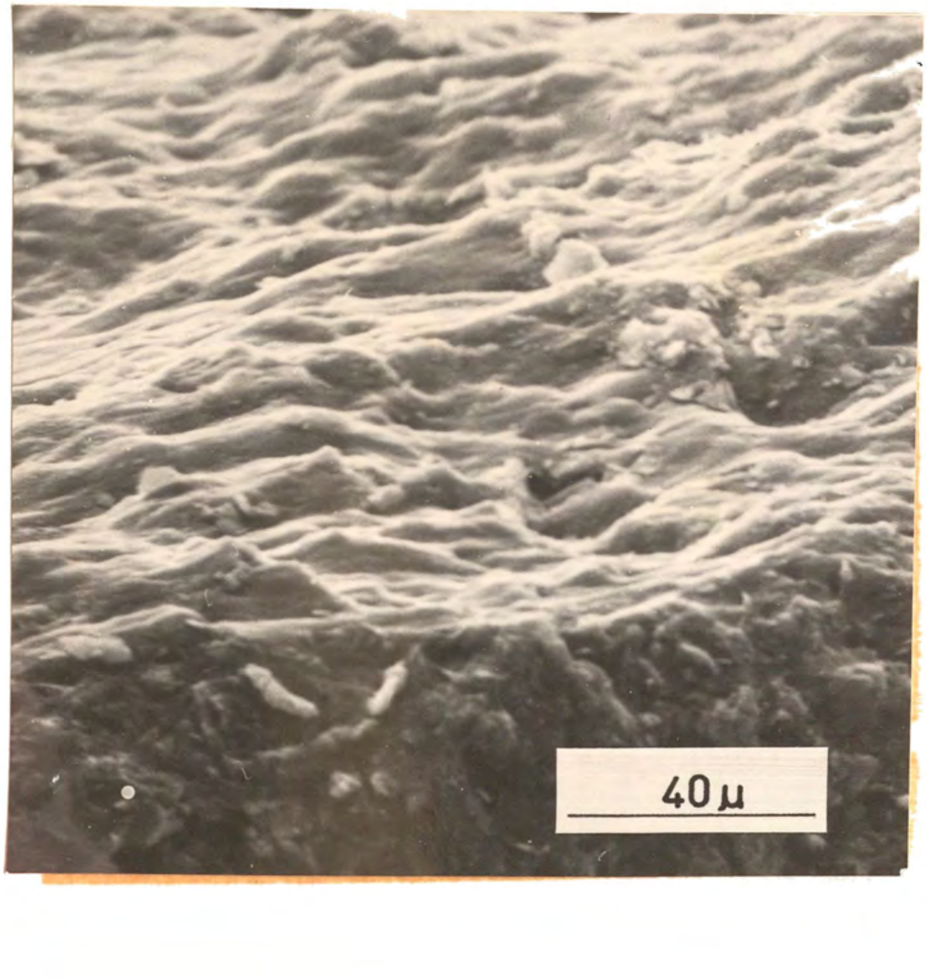

2µ

Plate 3.3. Transmission electron micrograph of the clay fraction (<2μ) of a Jurassic siltstone (10203). The clay particles include abundant illite and common montmorillonite and mixed-layer clay minerals.

Plate 3.4. Transmission electron micrograph of the clay fraction ($<2\mu$) of a Waikato Coal Measures mudstone (10208). Clay minerals consist of tubular halloysite (1), hexagonal kaolinite (2), and irregularly shaped illite (3).

Plate 3.5. Scanning electron micrograph of the fractured surface of a Waikato Coal Measures mudstone .(10210) in which the grain fabric is orientated parallel to the plane of bedding.


1
2
3
1μ


40μ

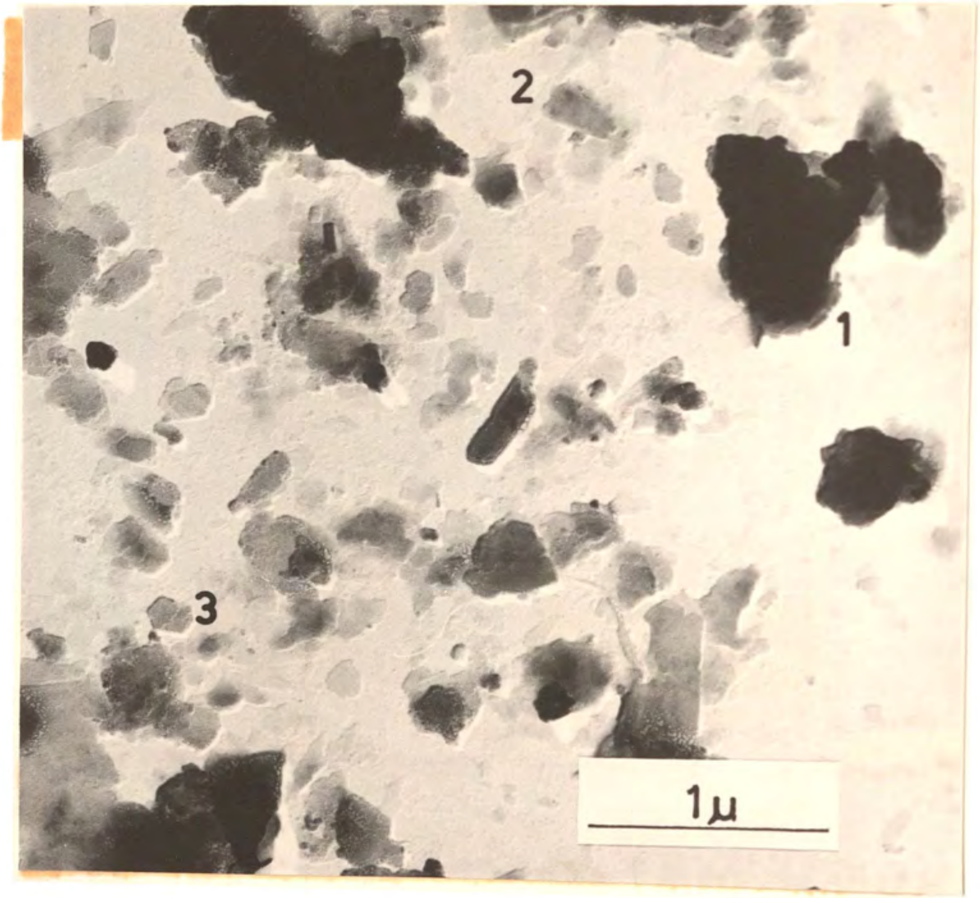
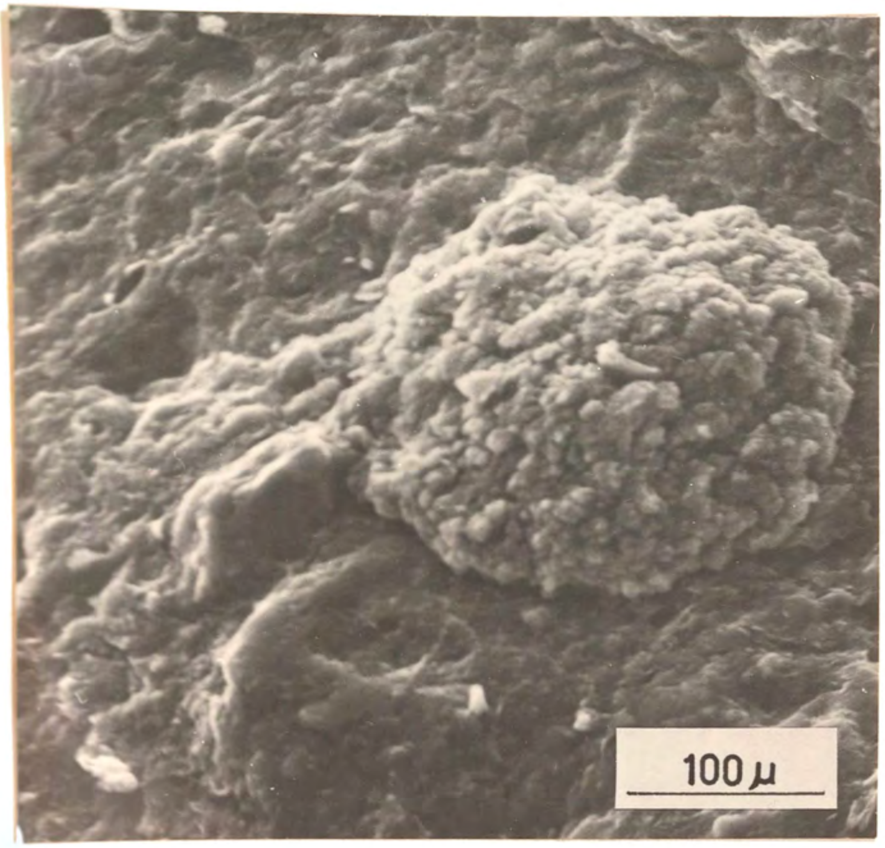

2
1
3
1µ

Plate 3.6. Transmission electron micrograph of the clay fraction ($<2\mu$) of a Mangakotuku Siltstone (10213). Note that the illite and chlorite (1) particles are larger than those of montmorillonite (2) and kaolinite (3).

Plate 3.8. Scanning electron micrograph showing a concretionary aggregate of montmorillonite in a sample of Mangakotuku Siltstone (10215) from the type section.

Plate 3.9. High resolution scanning electron micrograph of the concretionary aggregate shown in Plate 3.8 showing "waxy" flakes of montmorillonite.


100 µ

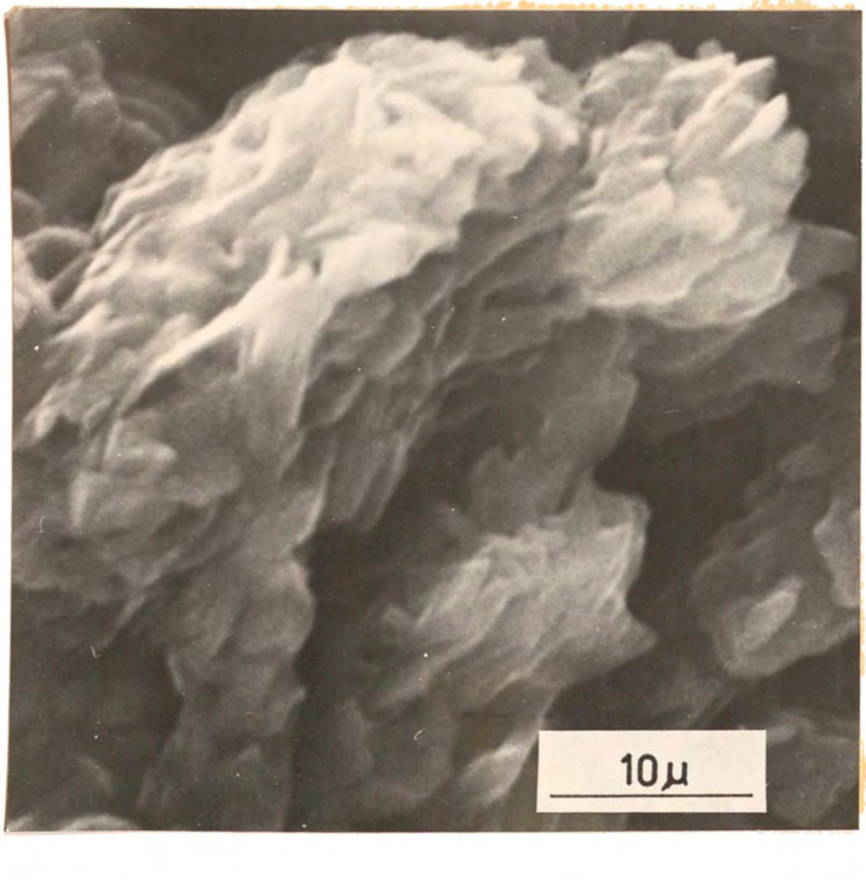
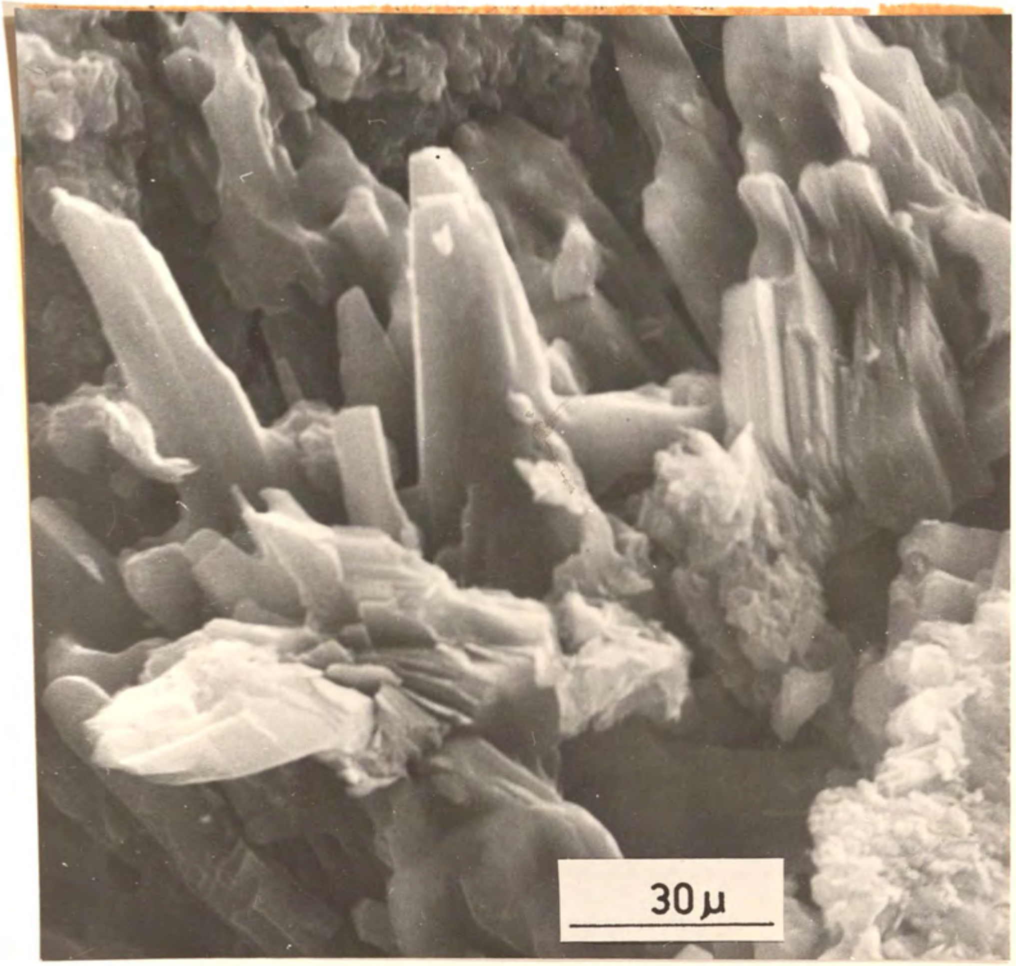
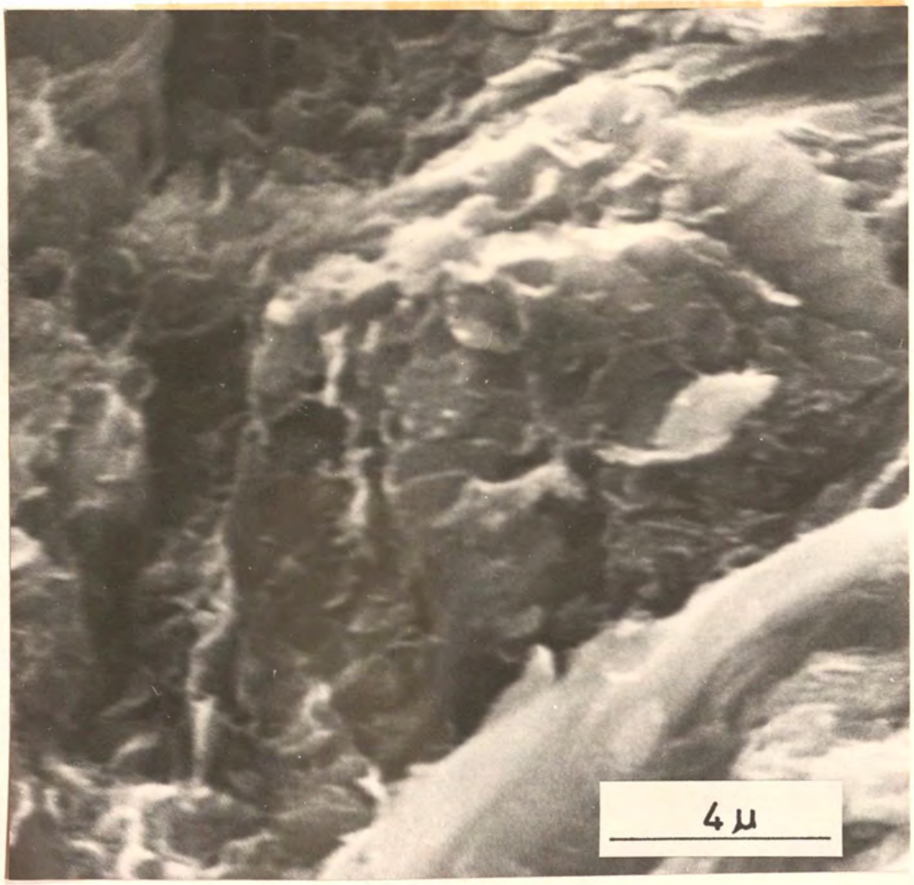

10 µ

Plate 3.7. Scanning electron micrograph of a sample (10213) of Mangakotuku Siltstone showing a zone of euhedral kaolinite crystals flanked by smaller "waxy" aggregates of montmorillonite.

Plate 3.10. Scanning electron micrograph of a fractured surface of Whaingaroa Siltstone (10222) from the type locality. Abundant thin flakes of montmorillonite coat the surfaces of quartz and feldspar grains and infill pore spaces.


30 µ


4 µ

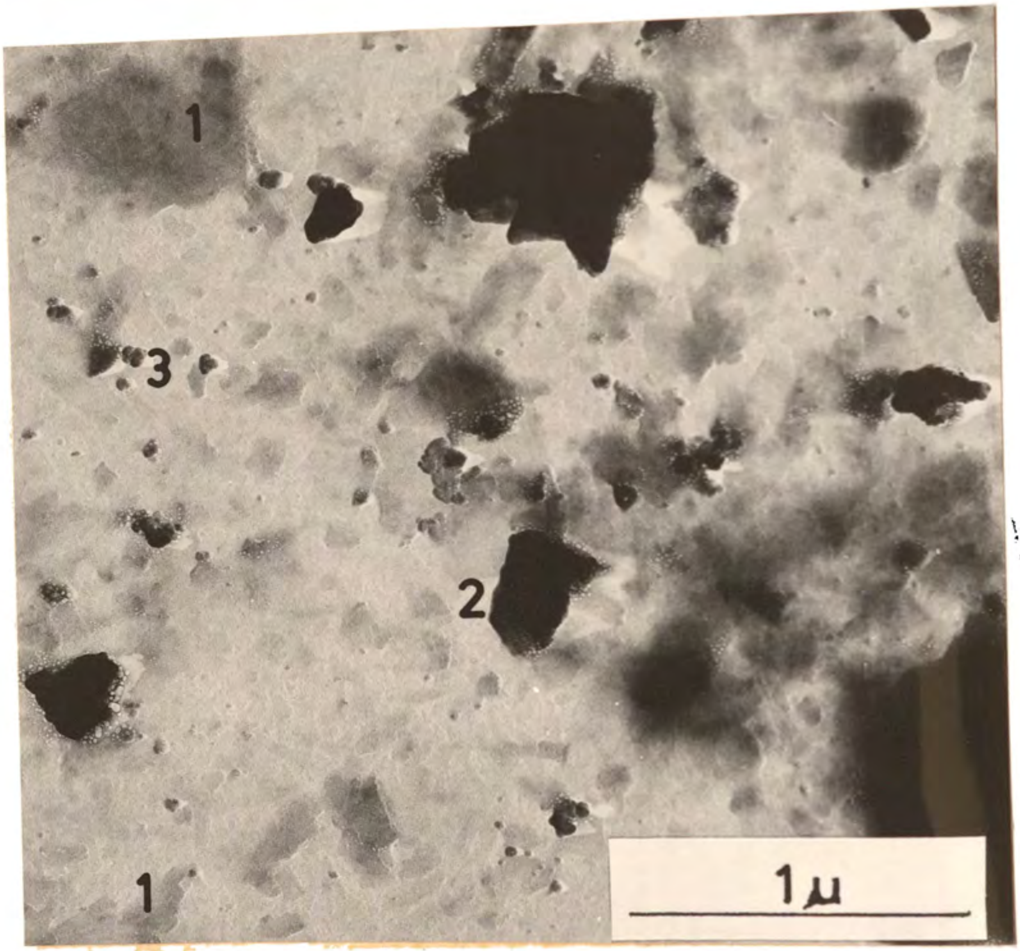
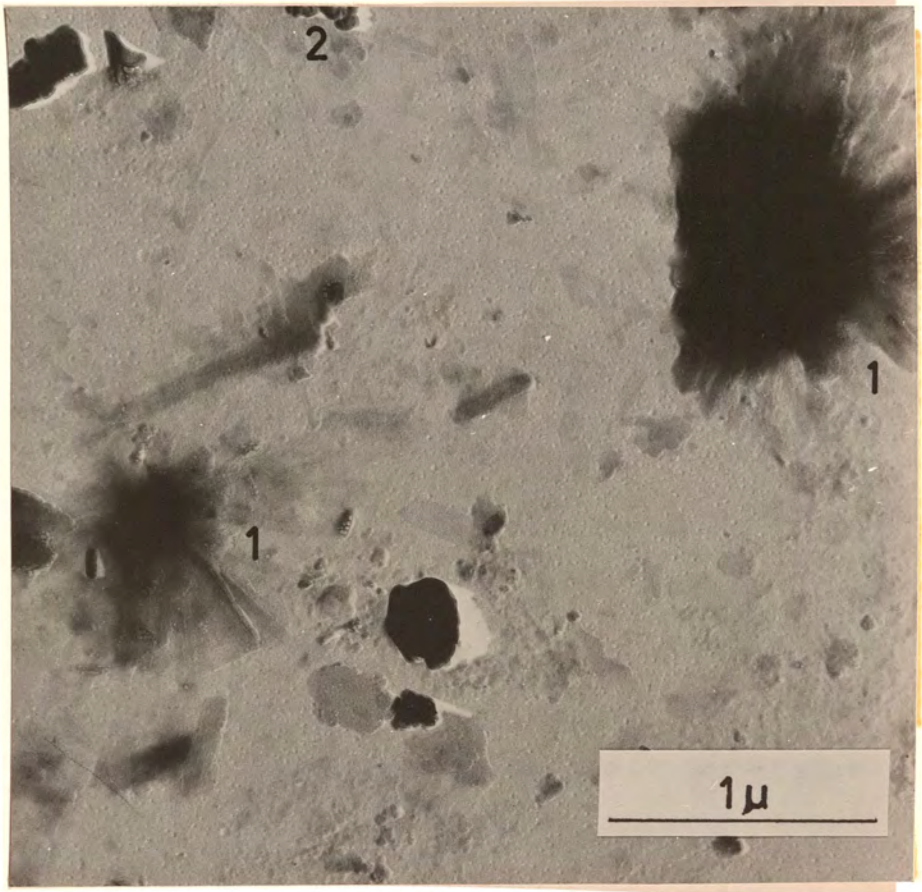

1
3
2
1
1µ

Plate 3.11. Transmission electron micrograph of the <2μ size fraction of an Aotea Sandstone sample (10227) from the type section. The thin lath-shaped units are montmorillonite (1), the larger, darker material illite (2), and XRD data suggests the tiny rounded "blebs" may be α-cristobalite (3).

Plate 3.12. Transmission electron micrograph of the clay fraction of a sample of Te Akatea Siltstone (10236) from the type section showing lath-shaped aggregates of montmorillonite (1), and small rounded "blebs" that may be α-cristobalite (2).

Plate 3.13. Transmission electron micrograph of clay material from the Whaingaroa Siltstone (10219) showing lath-shaped aggregates of montmorillonite (1), micaceous material (2), and small "blebs" of α-cristobalite (3). Much of the "shadowy" background material is montmorillonite.


2
1
1
1µ

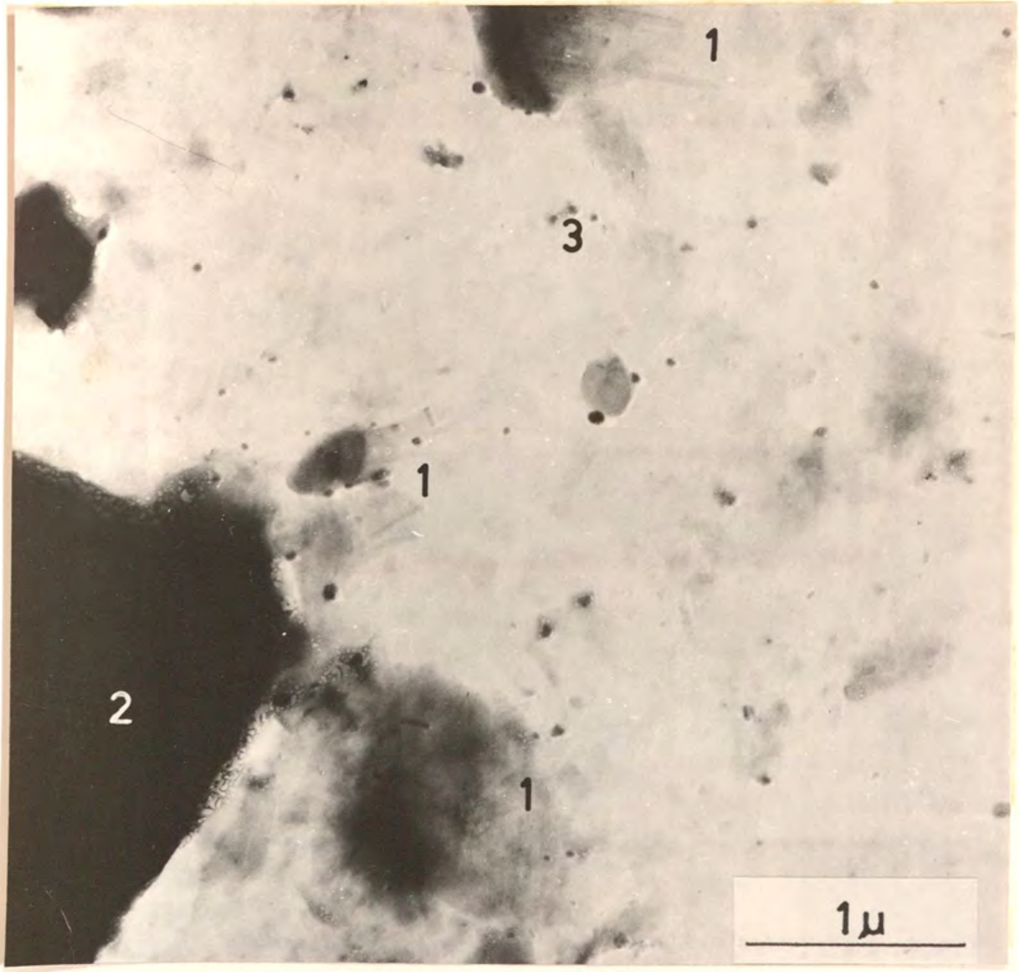
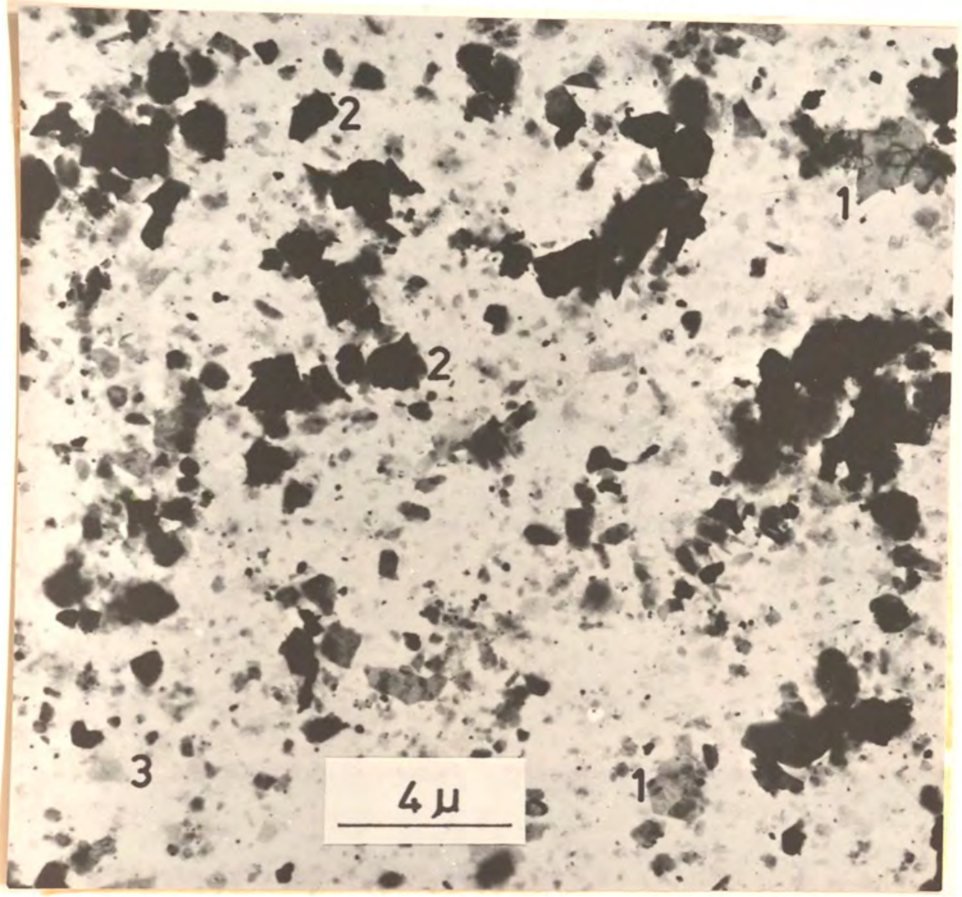
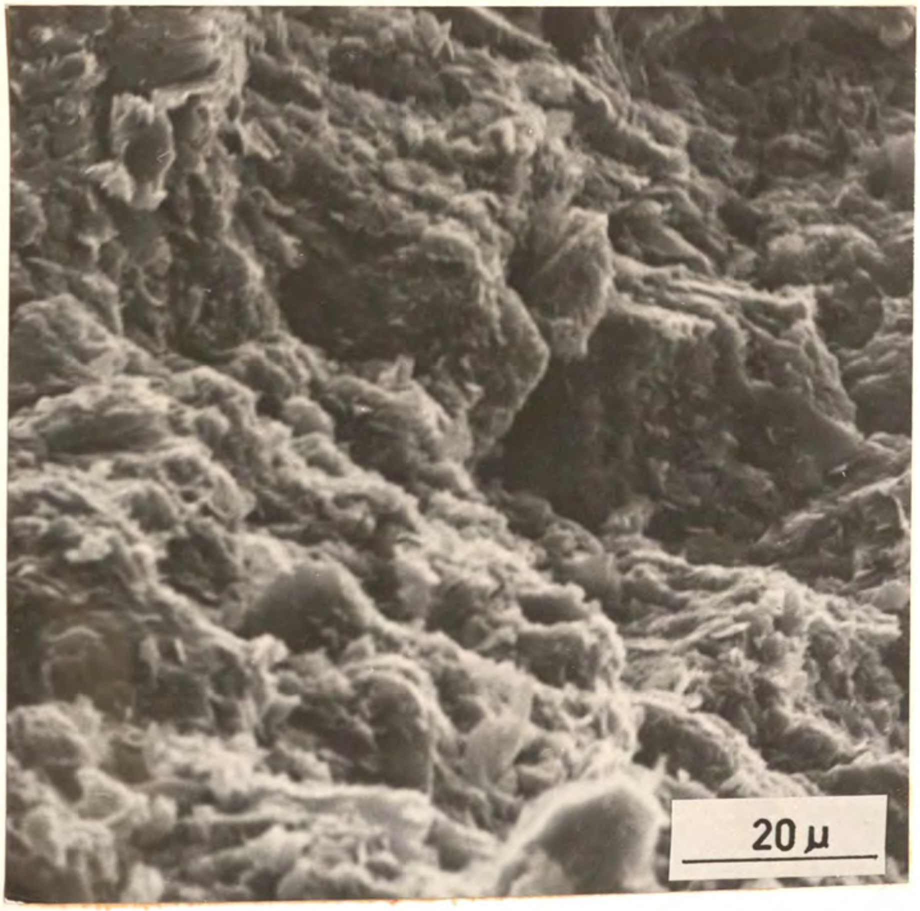
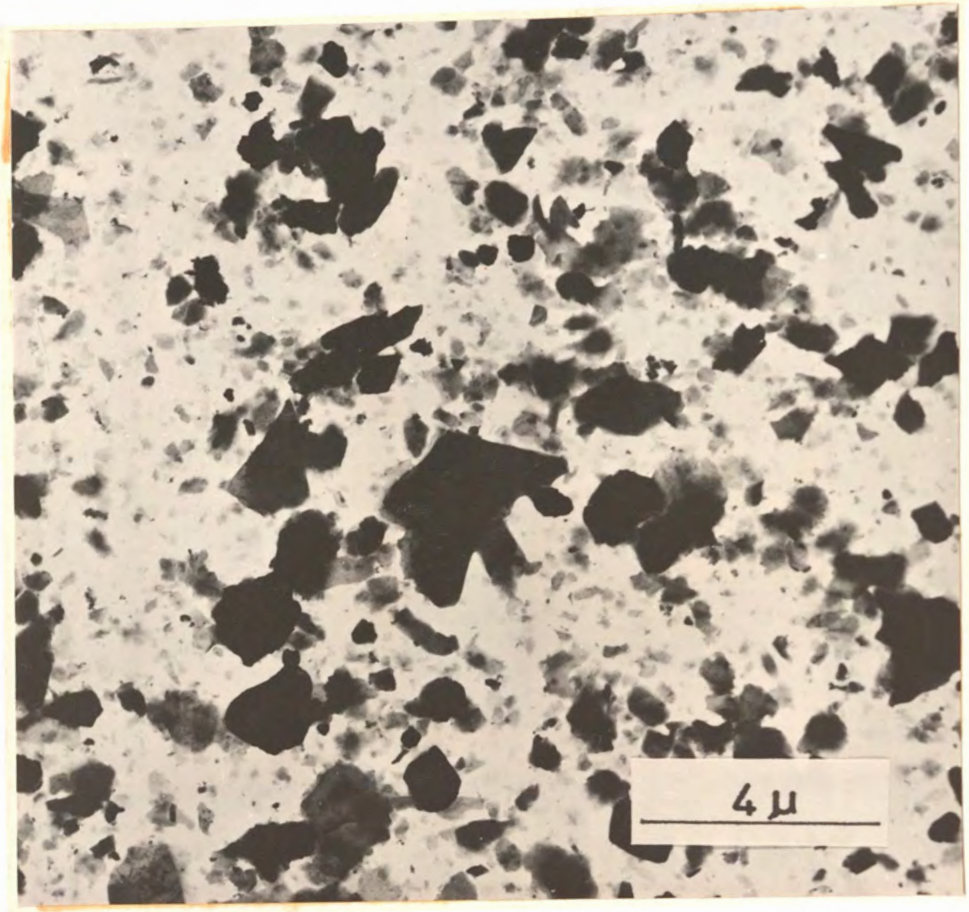

1
3
1
2
1
1µ

Plate 3.14. Transmission electron micrograph of the clay fraction of a Taumatamaire mudstone (10256). Note the significant increase in illitic (1) and chloritic (2) material compared to the montmorillonite-rich Oligocene mudstones (Plates 3.12 and 3.13). Montmorillonitic material (3) is characterised by diffuse grain boundaries.

Plate 3.15. Scanning electron micrograph of a fragment of Taumatamaire mudstone (10256). Clay and silt particles show a weakly orientated grain fabric running from top left to bottom right of the figure.


2
1
2
3
1
4 µ


20 µ


4 µ

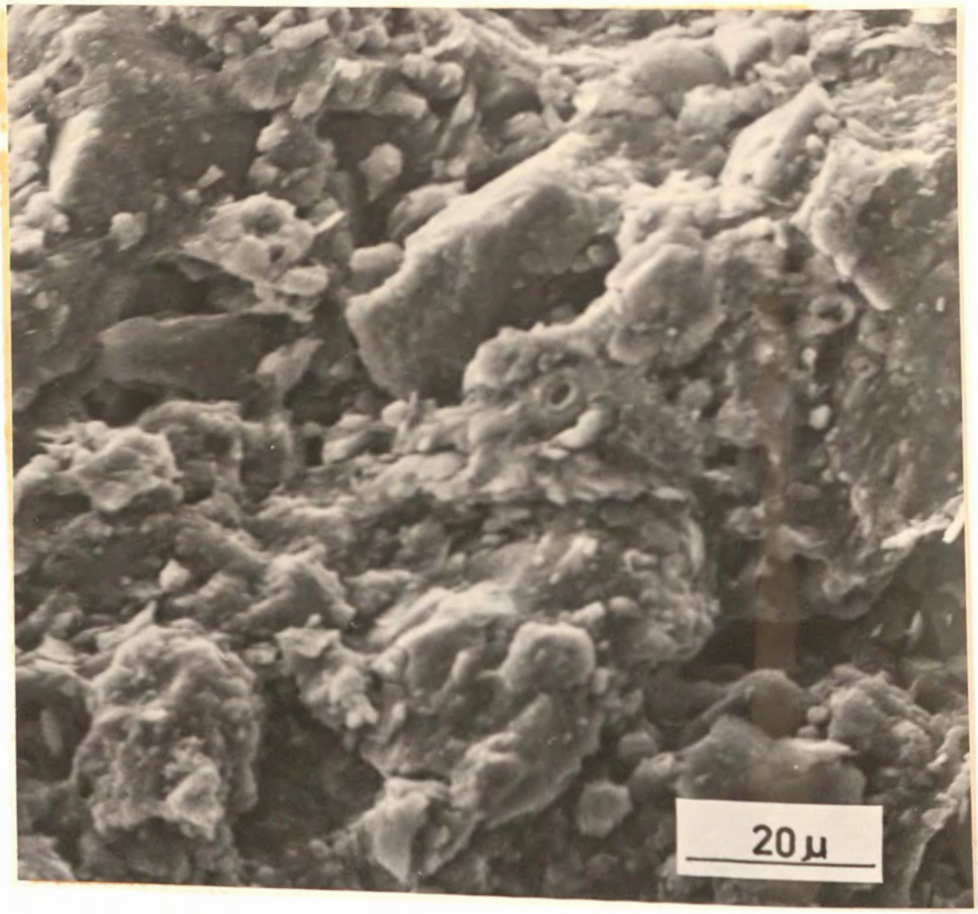
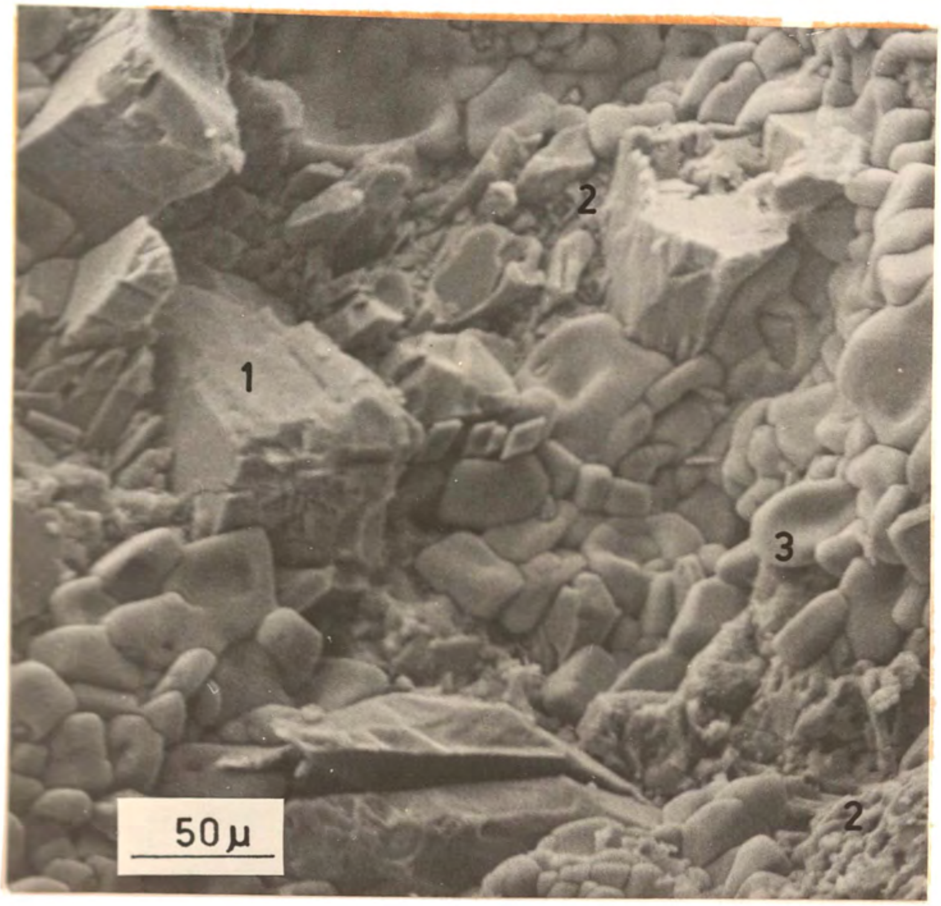

20 µ

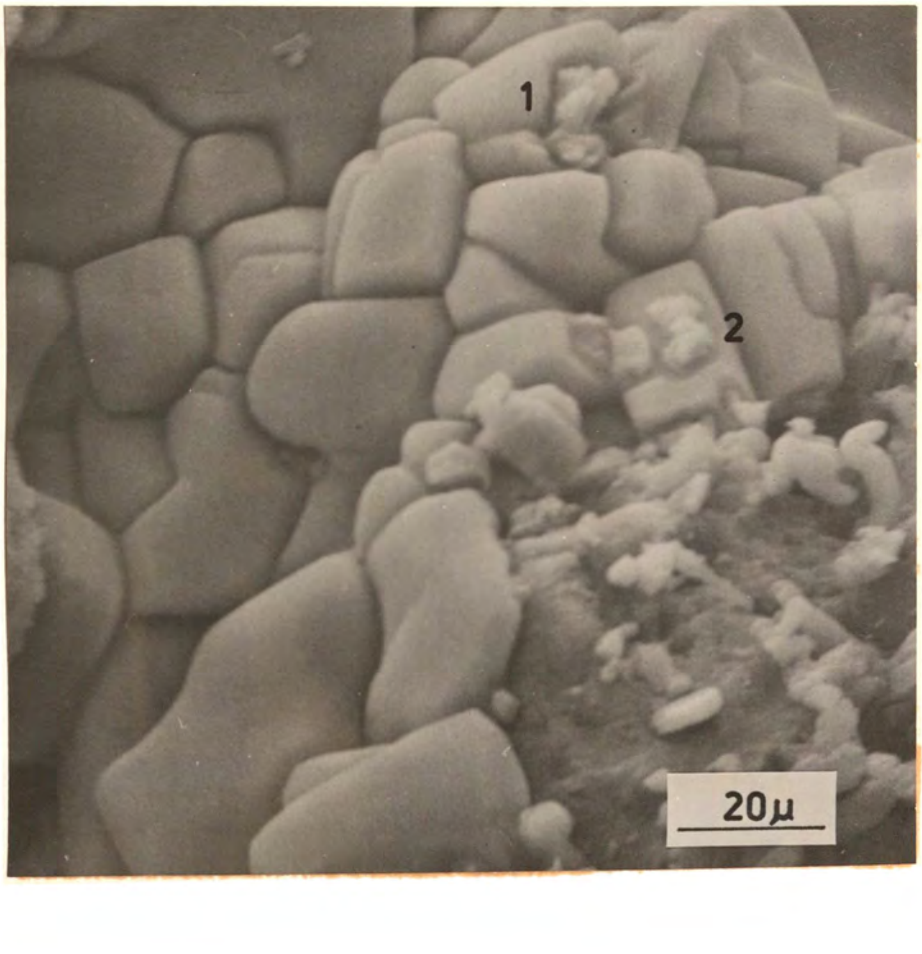
Plate 3.16. Transmission electron micrograph of clay fraction material from the Lower Mokau Sandstone (10279). Note the predominance of angular, detrital particles of illite and chlorite. Montmorillonite constitutes the fine-grained background material.

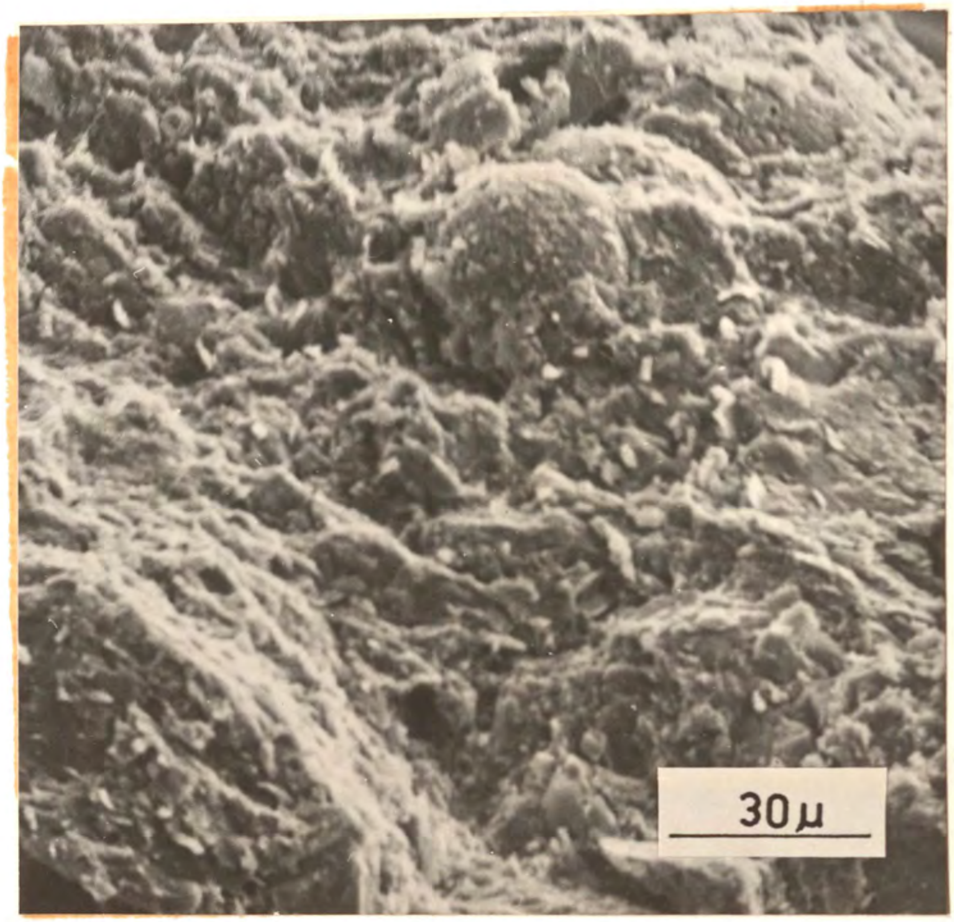
Plate 3.17. Scanning electron micrograph of a fractured surface of a sample of Upper Mokau mudstone (10287) displayin a rather unorientated porous grain fabric. Clay material occupies the interstitial spaces between larger quartz and feldspar grains.

Plate 3.18. Scanning electron micrograph of the tuffaceous Purupuru Volcanic Sandstone (10297) showing large laths of plagioclase feldspar (1) in a matrix of montmorillonite (2) and α-cristobalite (3). The interlocking fabric of the feldspar and cristobalite material suggest a diagenetic origin for the cristobalite.

Plate 3.19. A higher resolution scanning electron micrograph of part of the tuffaceous sandstone (10297) shown in Plate 3.18. Note how the cristobalitic material envelops clay material at (1) and (2) suggesting it formed subsequent to the montmorillonite.


2
1
3
2
50μ


1
2
20μ


30µ

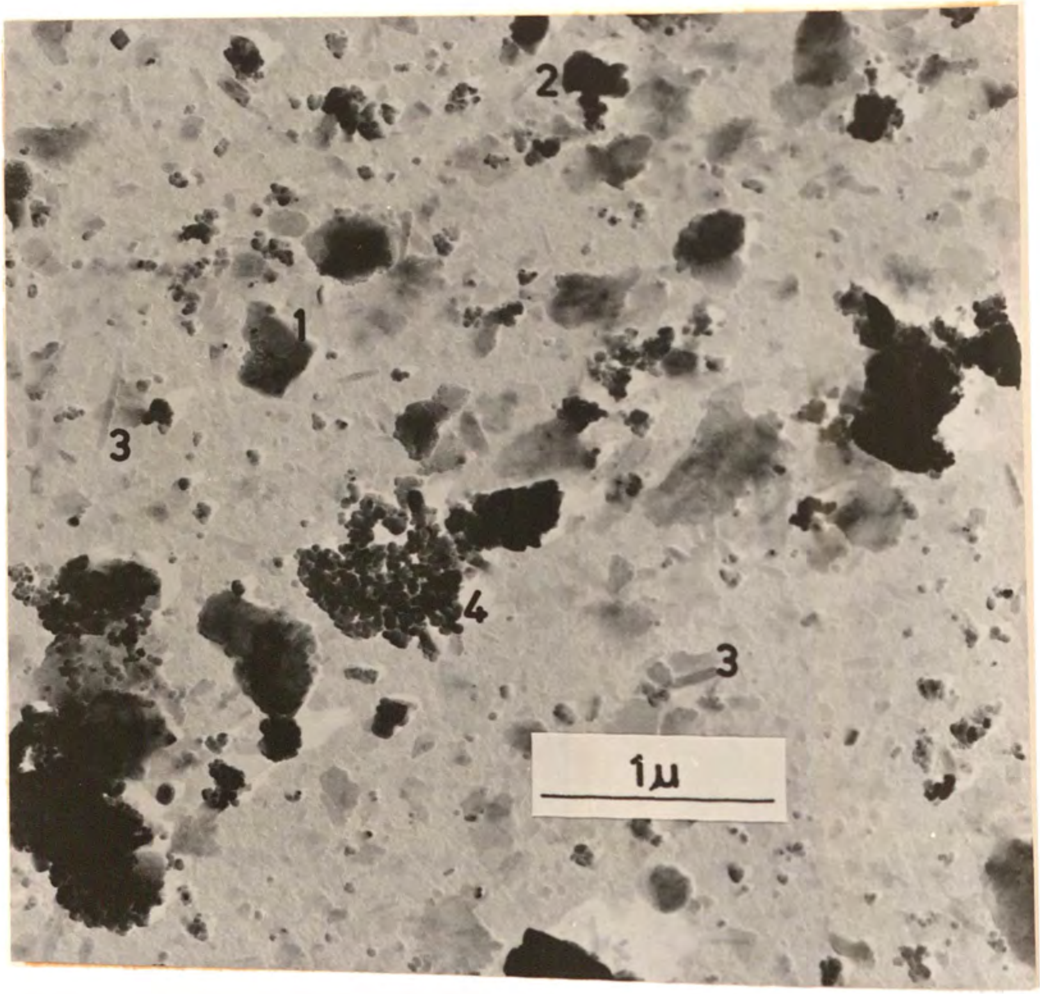
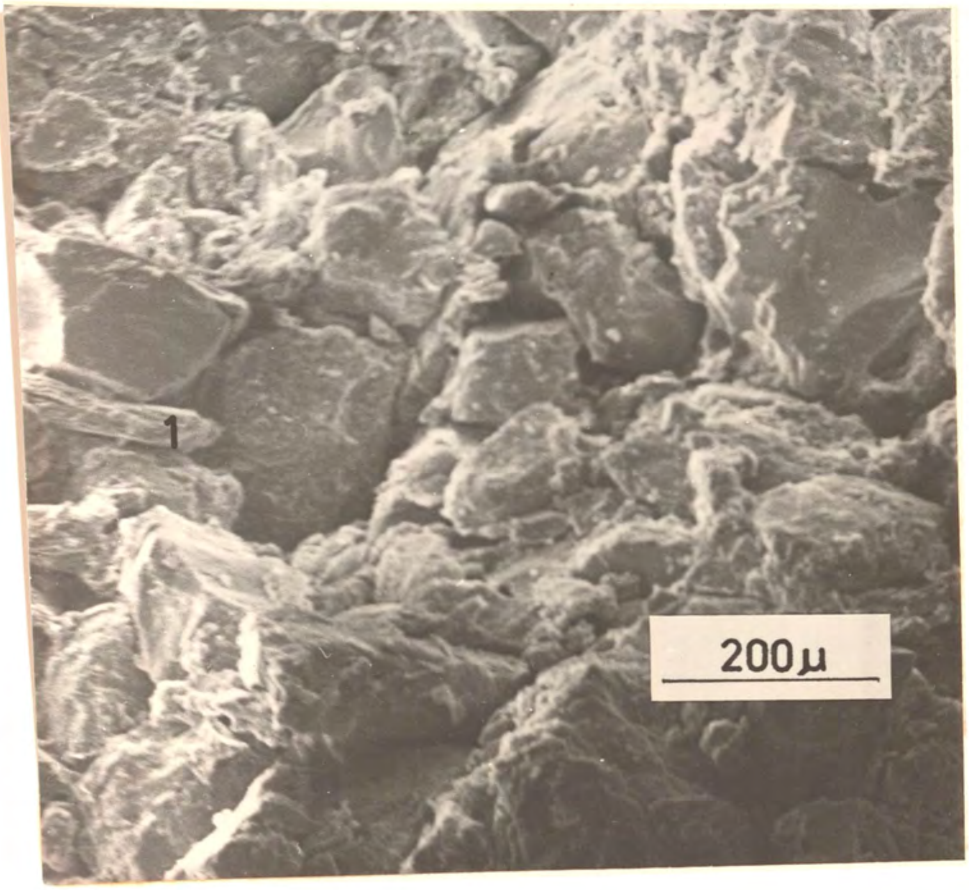

2
1
3
4
3
1µ

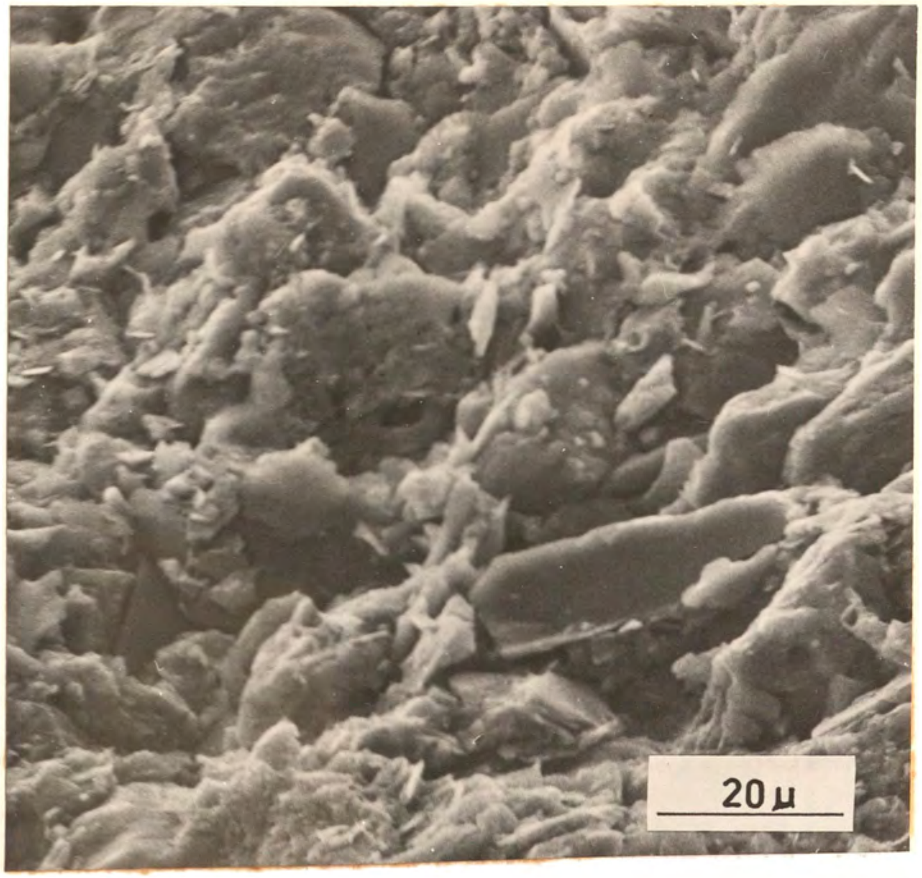
Plate 3.20 Scanning electron micrograph of the fractured surface of a Purupuru Volcanic mudstone sample (10296) from the type section showing a poorly orientated grain fabric rich in clay material. Note the occurrence of spherules above the centre of the figure; higher resolution electron micrographs indicate they consist of lath-shaped particles (probably montmorillonite) averaging $<2\mu$ in diameter (cf. Plates 3.8 and 3.9).

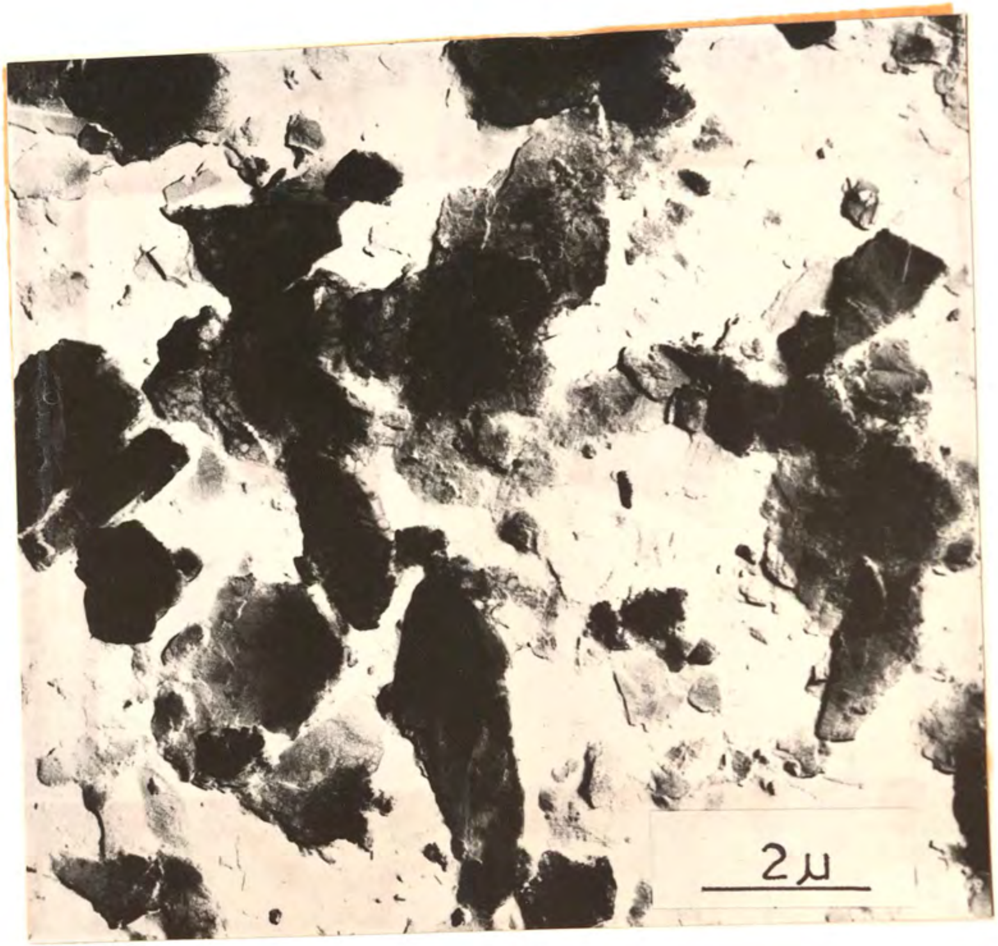
Plate 3.21. Transmission electron micrograph of the clay fraction from a tuffaceous Purupuru Volcanic mudstone (10296). Large illitic (1) and chloritic (2) particles are surrounded by low relief laths of montmorillonite (3) and small (0.03μ diameter) rounded "blebs" of α-cristobalite (4). Note the similarity of the montmorillonite laths and the cristobalite particles to those in the Oligocene sediments (Plates 3.11, 3.12 and 3.13).

Plate 3.22. Scanning electron micrograph of the fractured surface of a Ferry sandstone sample (10811) from the type section. This micrograph and other higher resolution micrographs show a porous unorientated grain fabric in which clay minerals exist as clumps of "books" (1) of particles between the larger, more equidimensional quartz, feldspar and rock-fragment grains.

Plate 3.23. Scanning electron micrograph of the fractured surface of a Ferry mudstone sample (10307) from the type section. Compared to the sandstone sample (Plate 3.22) there is a tighter, but still unorientated, grain fabric.


1
200µ


20µ


2 µ

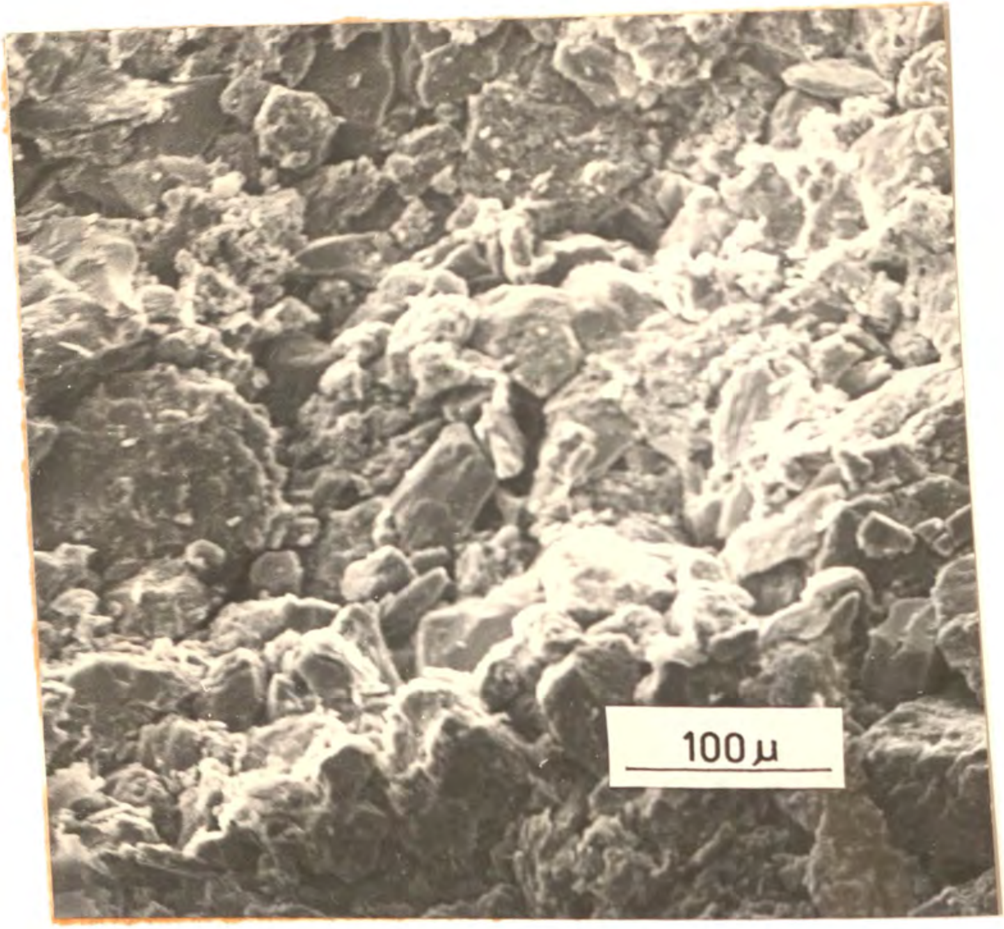
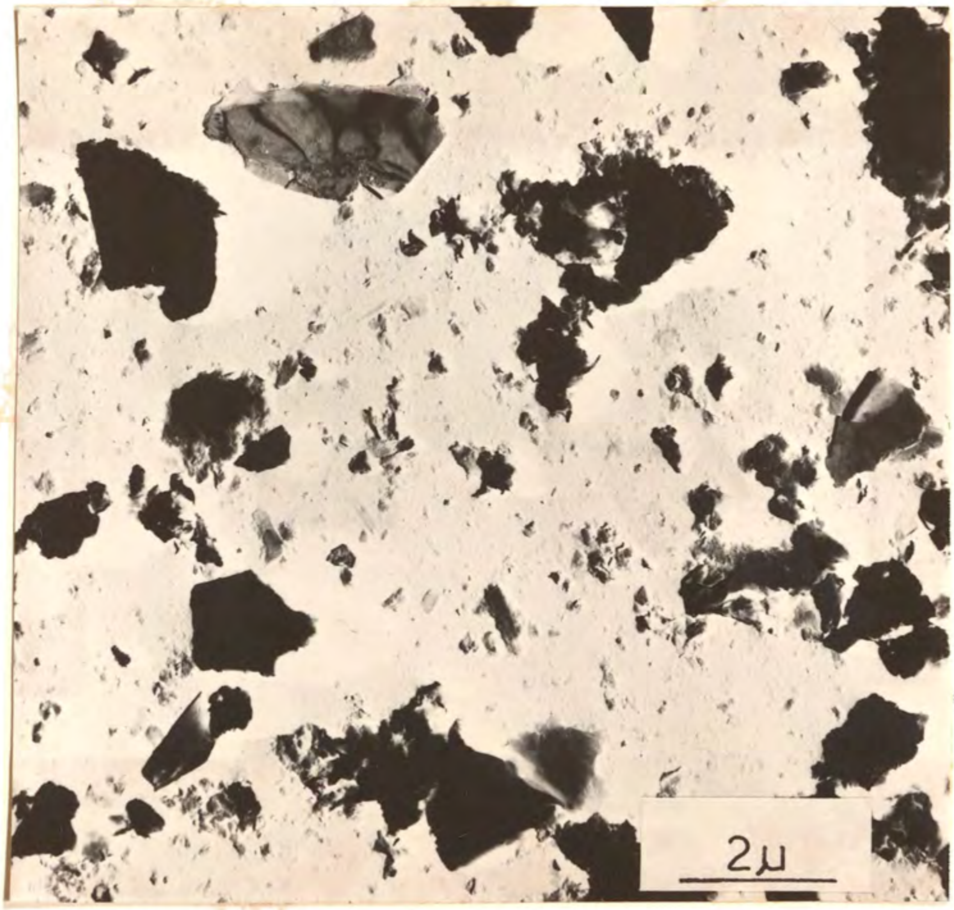
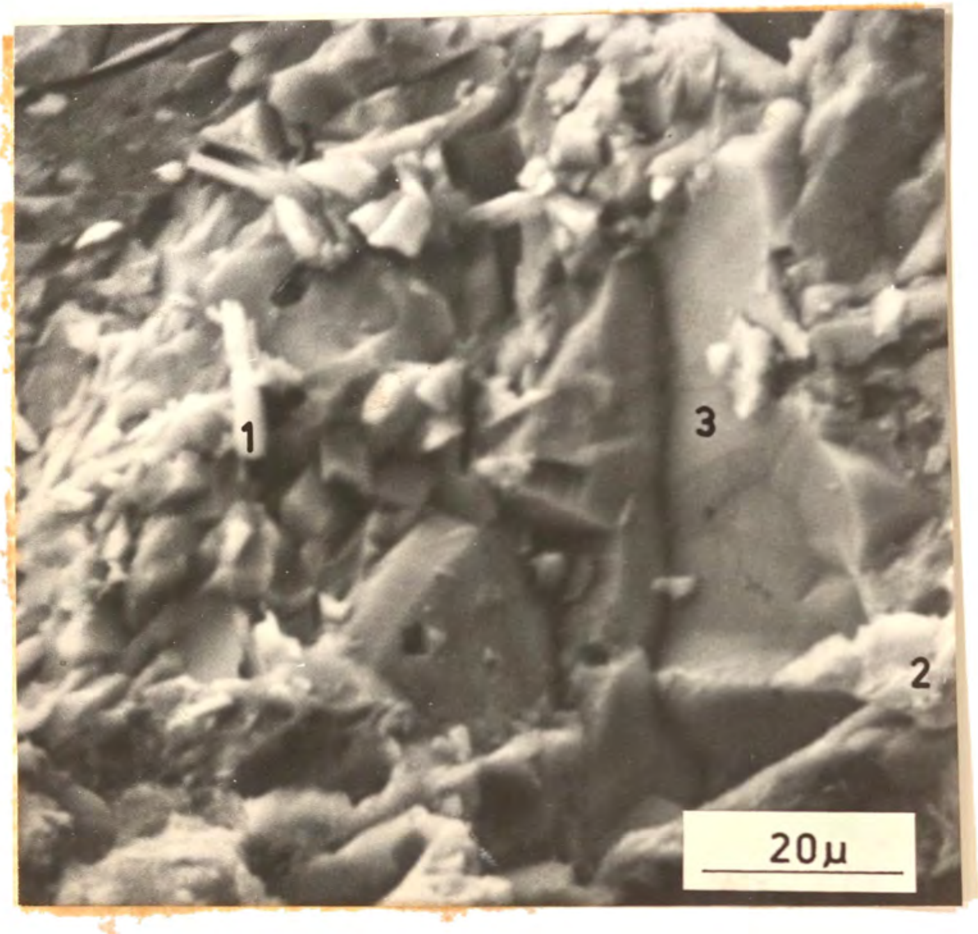

100 µ

Plate 3.24. Transmission electron micrograph of the clay fraction ($< 2\mu$) of a Mt. Messenger Sandstone sample (10317) showing abundant illite and common chlorite. Note the anhedral to subhedral form of the clay minerals.

Plate 3.25. Scanning electron micrograph of the fractured surface of a Mt. Messenger sandstone sample (10317) showing a porous random grain fabric. Clay material (largely illite and chlorite) occurs as silt-sized clumps between the larger terrigenous grains.

Plate 3.26. Transmission electron micrograph of the clay fraction ($<2\mu$) of a Urenui tuffaceous sandstone sample (10327) showing large particles of illite and chlorite, fine-grained particles and clumps of montmorillonite (centre and left of centre). Note the large illite particle (top left) showing wavy patterns produced during weathering.


2µ


1
3
2
20µ

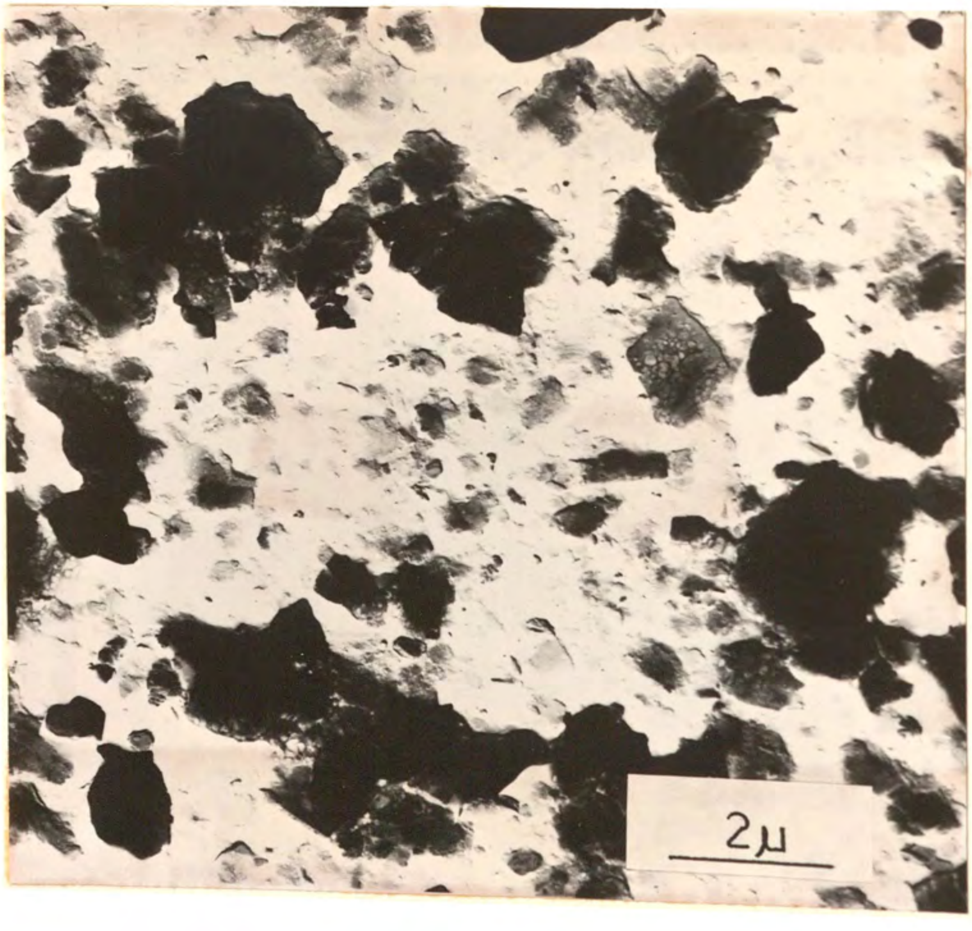
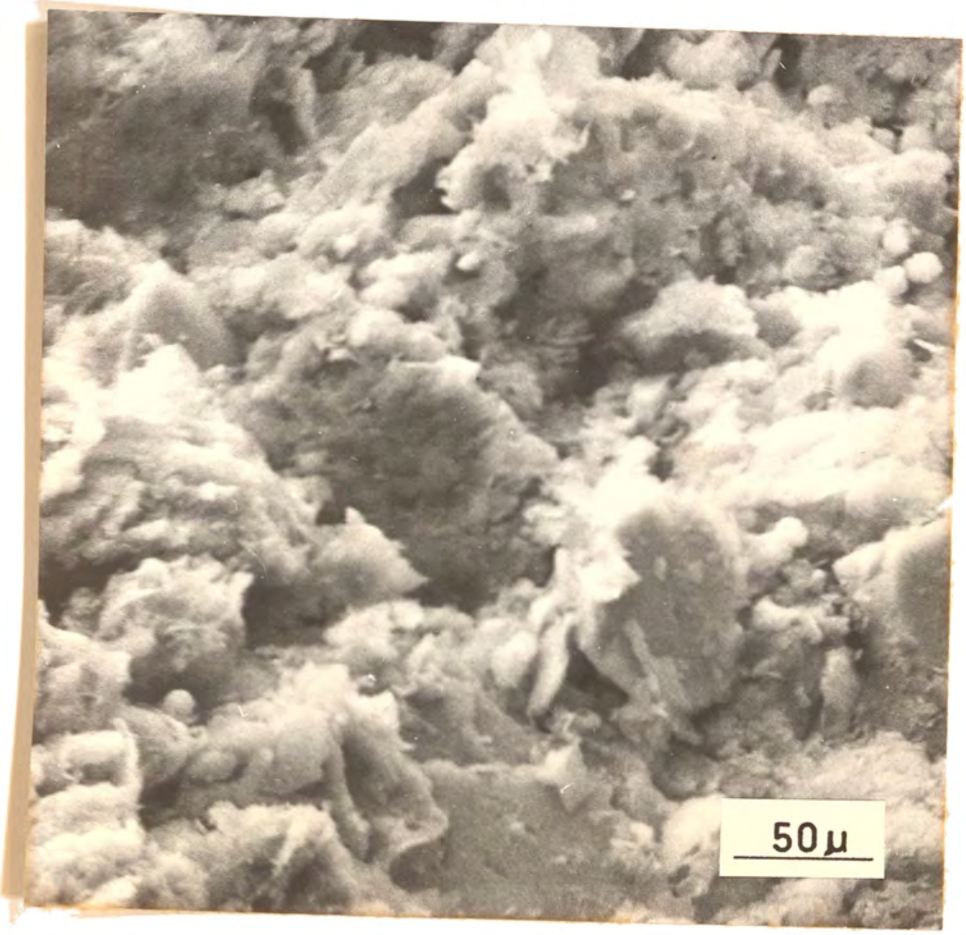

2µ

Plate 3.27. Scanning electron micrograph of the fractured surface of a tuffaceous Urenui sandstone (10327) showing clumps (2) and rod-like particles (1) of montmorillonite set in a matrix of large feldspar crystals and analcite (3). The clay material has a similar morphology to that in the tuffaceous Purupuru sandstone (Plates 3.18 and 3.19).

Plate 3.28. Transmission electron micrograph of the clay fraction ($<2\mu$) of a Urenui Siltstone sample (10324) from the type section. Note the anhedral to subhedral form of the illite and chlorite particles and the "bubbly" patterns on some of the grains (e.g. grain to right of centre).

Plate 3.29. Scanning electron micrograph of a fractured surface of a Urenui siltstone (10324) from the type section. The clay material is mostly illite and chlorite.

Plate 3.30. Scanning electron micrograph of a fractured surface of a sample of Matemateaonga Sandstone (10334) from the type section area. Note the random grain fabric, and the large amount of clay material coating the sand grains and filling the interstitial pores.


50µ

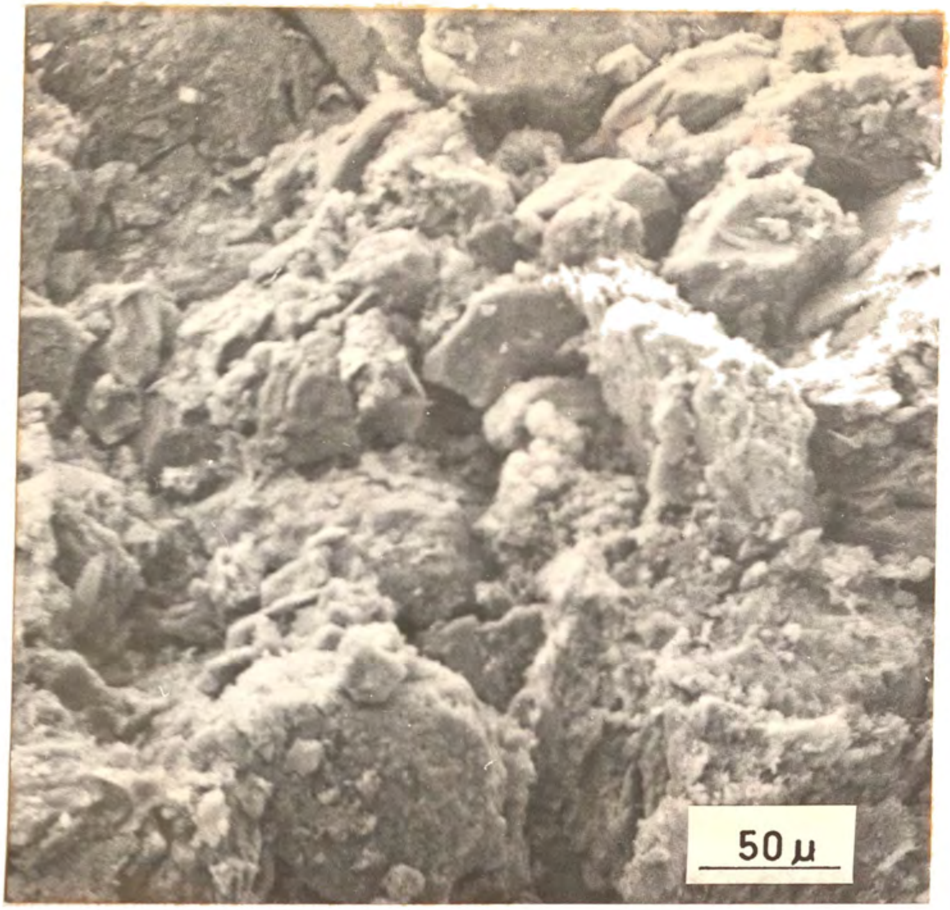
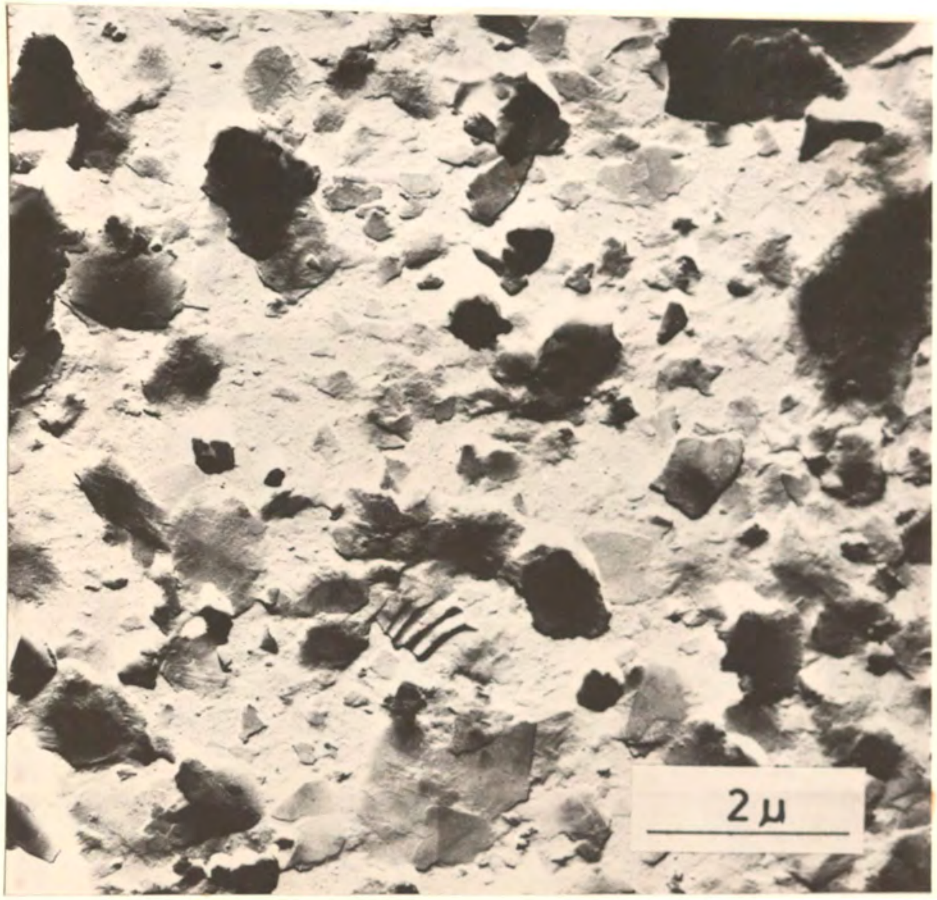

50µ

Plate 3.31. Transmission electron micrograph of the clay fraction from the Matemateaonga Sandstone (10334) showing an abundance of detrital illite and chlorite.

Plate 3.32. Scanning electron micrograph of a fractured surface of the Tangahoe Mudstone (10341) from the type section. The random grain fabric is characterised by the presence of silt-sized laths of illite and chlorite (1). Higher resolution micrographs suggest that montmorillonite constitutes the smaller grain aggregates (2).


2 µ

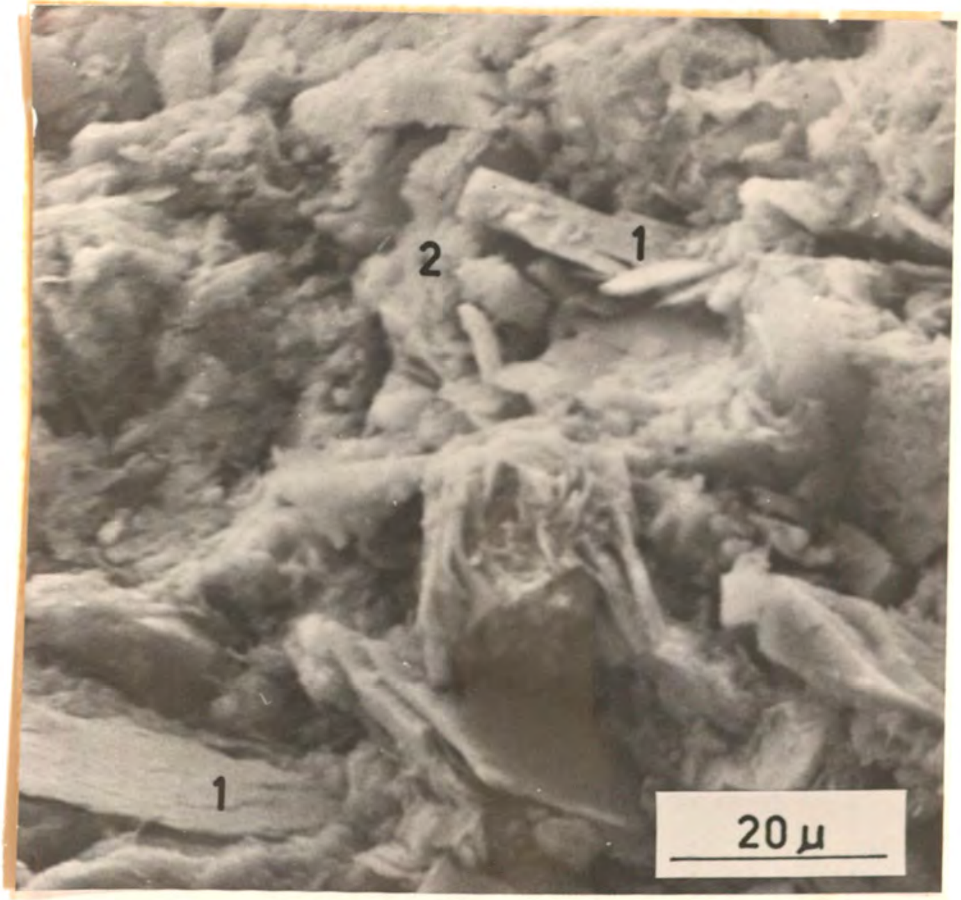

1
2
1
20 µ

Plate 3.33. Scanning electron micrograph of a fractured surface of a Wanganui mudstone sample (10354) showing the poorly sorted, non-porous, clayey nature of the sediment. A crude orientation of grain fabric is evident from bottom left to top right.

Plate 3.34. Transmission electron micrograph of a Hawera Series silt (10359) showing fine-grained montmorillonite (1), large laths of illite (2), and small rounded particles of α-cristobalite (3).


20µ

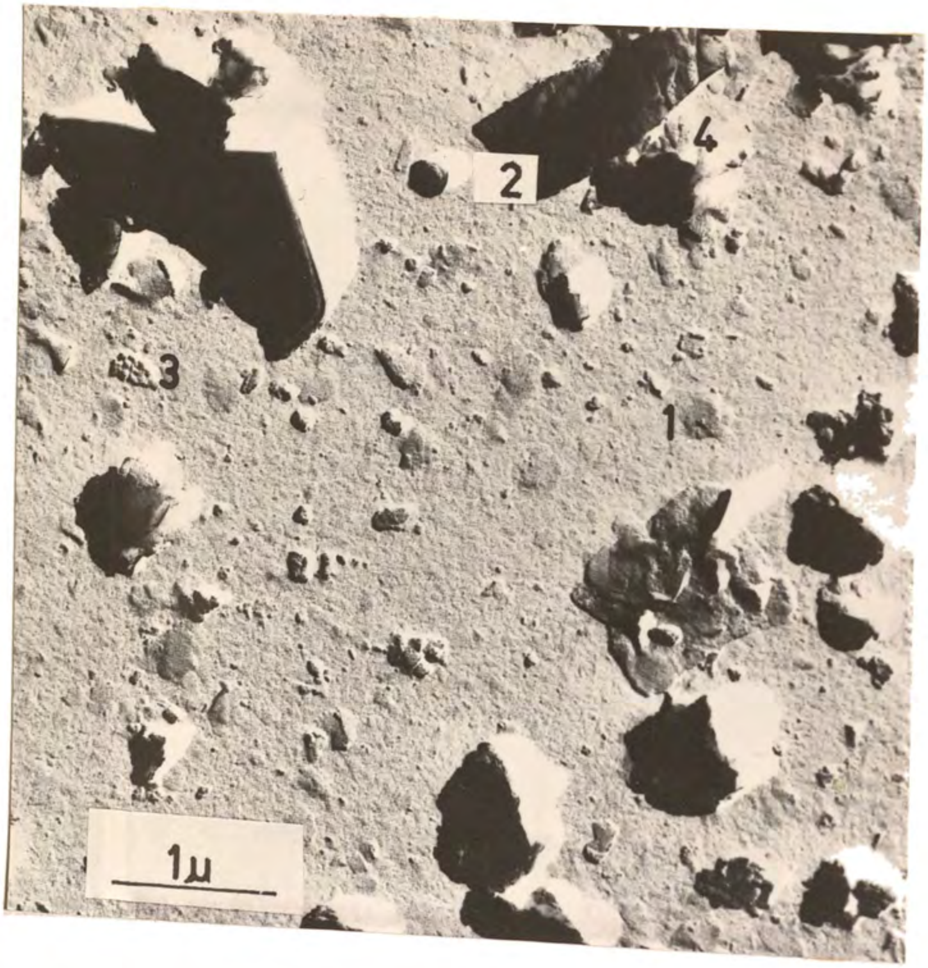
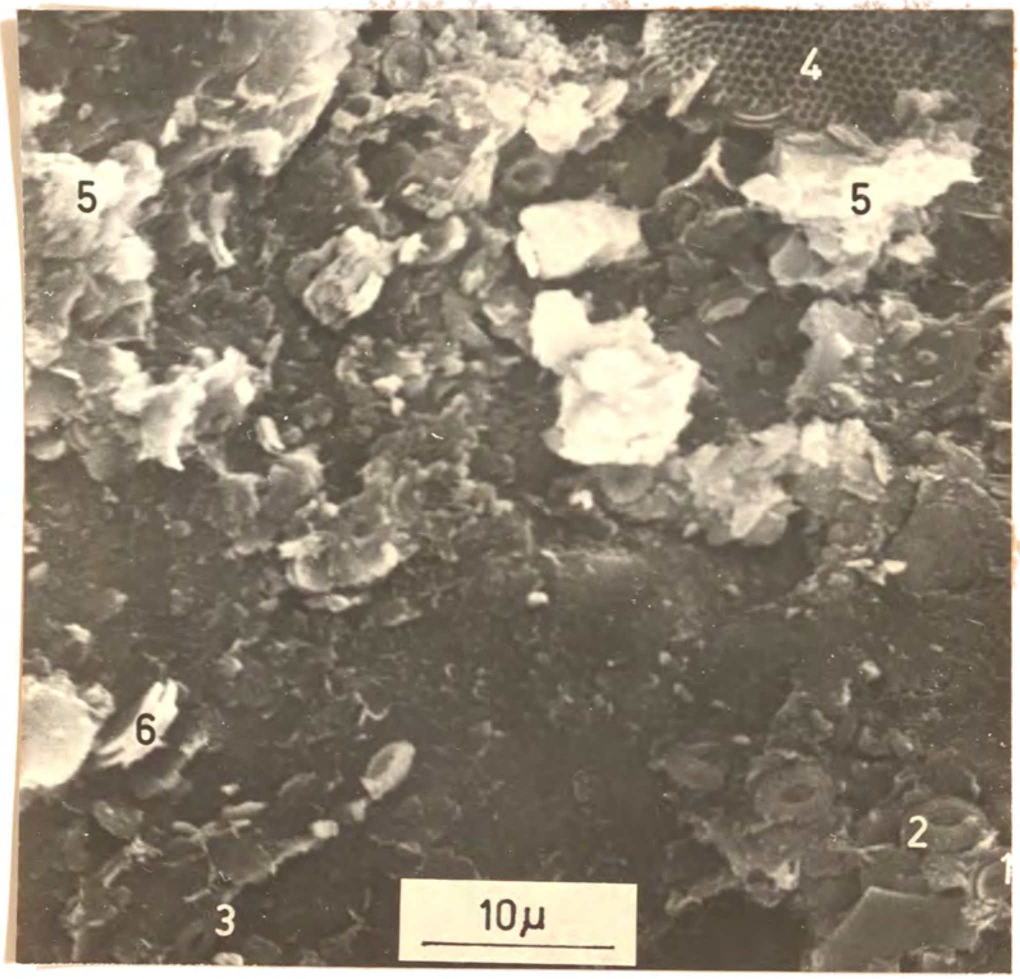

2
4
3
1
1µ

Plate 4.1. Scanning electron micrograph of a Hamilton Shelf surficial sediment sample (B778). Biogenic material identified consists of coccoliths: *Helicopontosphaera kamptneri* (1), *Cyclococcolithus leptoporous* (2), *Emiliania huxleyi* (3), and diatom fragments (4). Clay material is identified as montmorillonite (5) and kaolinite (6).

Plate 4.2. Transmission electron micrograph of typical fine-grained ($<2\mu$) material from a Hamilton Shelf surficial sediment sample (B780). Clay minerals consist mainly of low relief, fine-grained, montmorillonite particles (1), anhedral illitic (2) and chloritic (3) material, and some kaolinite (4). Much of the fine-grained background material may represent amorphous siliceous material.


4
5
5
6
2
1
3
10µ

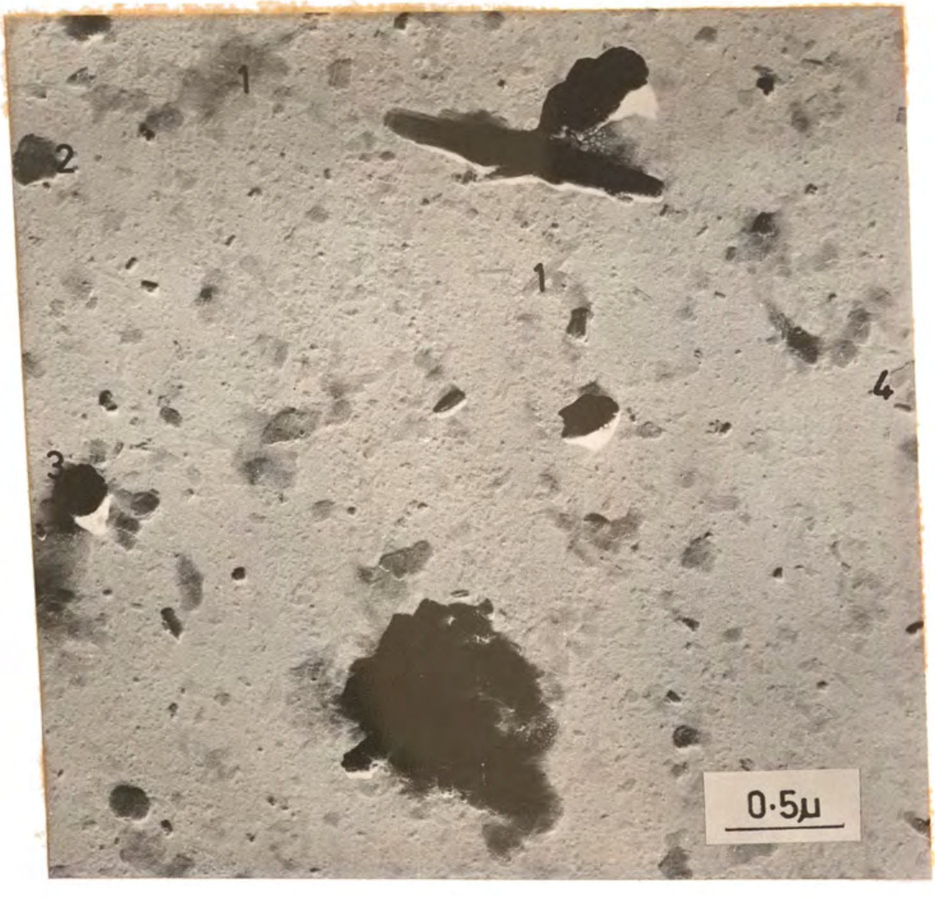
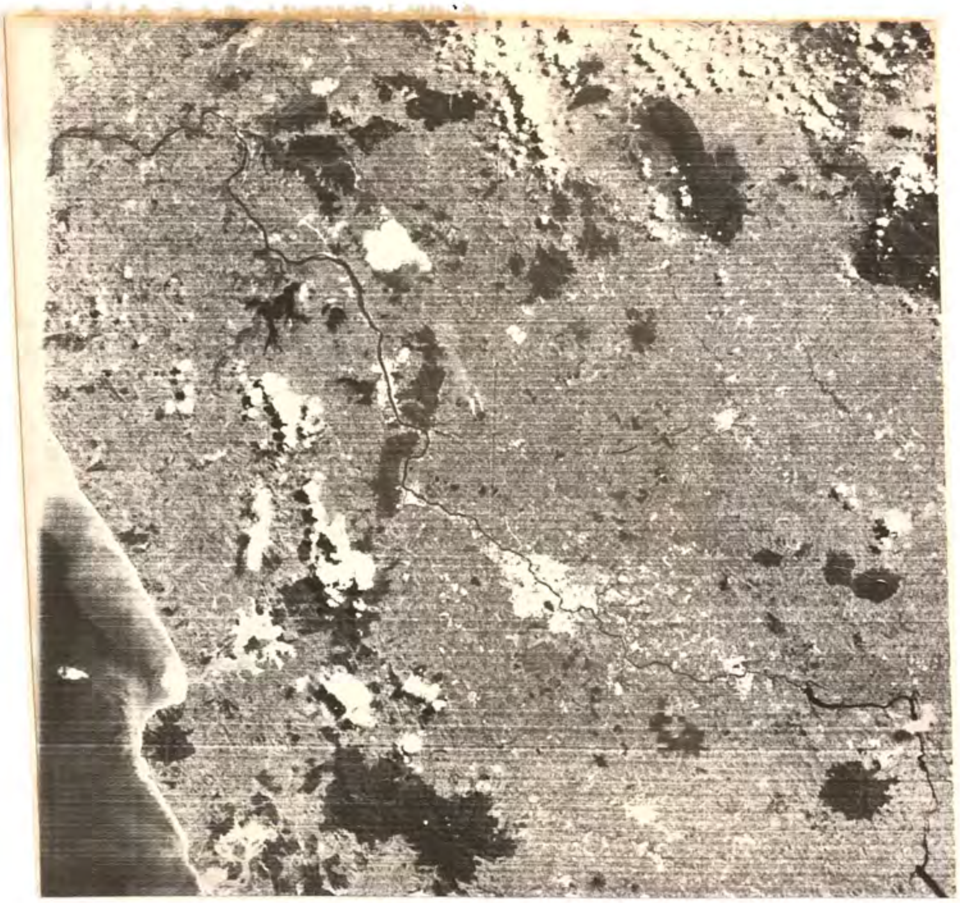

1
2
1
4
3
0·5µ

Plate 4.3. Exposure of Whaingaroa Siltstone on the shores of Raglan Harbour (N64/473481). Note the vertical runnels cut in the cliff face by the action of surface run-off on the readily erodible siltstone (cf. Plate 1.4).

Plate 4.4. LANDSAT image (16 February, 1976; Band 4) of the Manukau Harbour and adjacent coastal areas. Note that suspended sediment emerging from the Manukau Harbour shows movement to the north while suspended sediment from the Kaipara Harbour drifts south, directions compatible with those of major ocean currents (cf. Fig. 4.2). Scale 1 cm : 12 km.

Plate 4.5. LANDSAT image (15 February, 1976; Band 4) showing suspended sediment emerging from the Raglan and Aotea Harbours some 10 km onto the shelf. Scale 1 cm : 9 km.

Plate 4.6. LANDSAT image (30 October, 1975; Band 4) showing suspended sediment emerging from the Raglan and Aotea Harbours and drifting to the north. Scale 1 cm : 9 km.



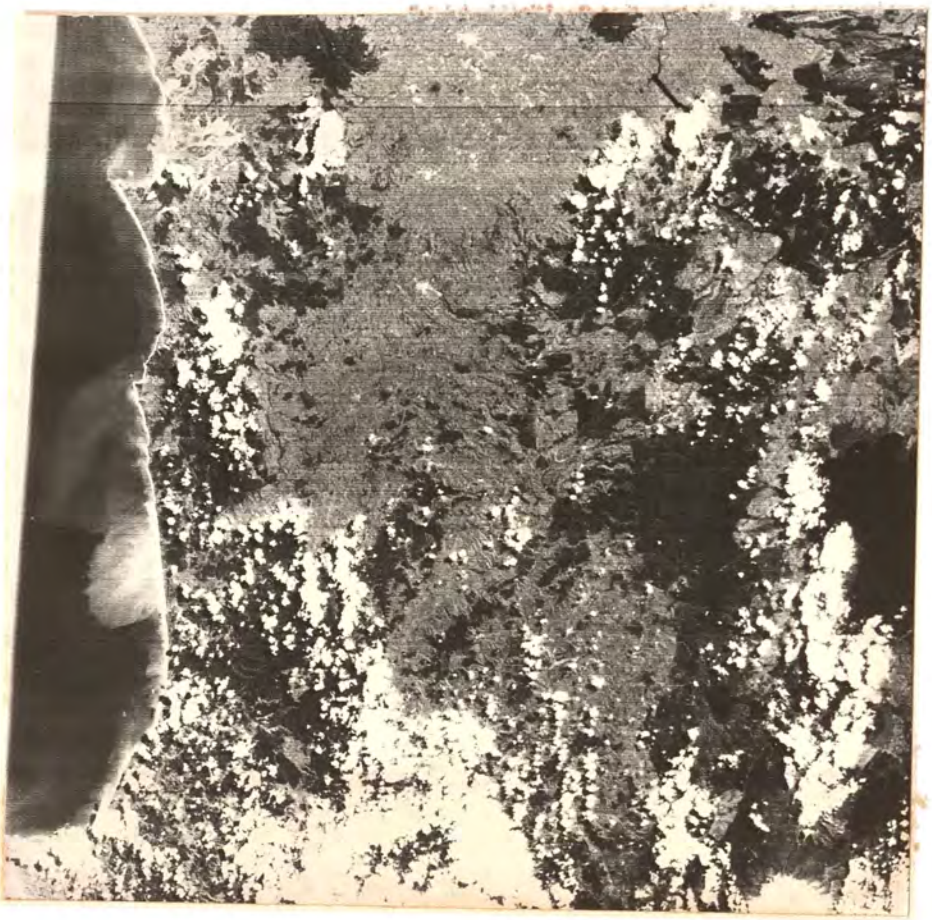


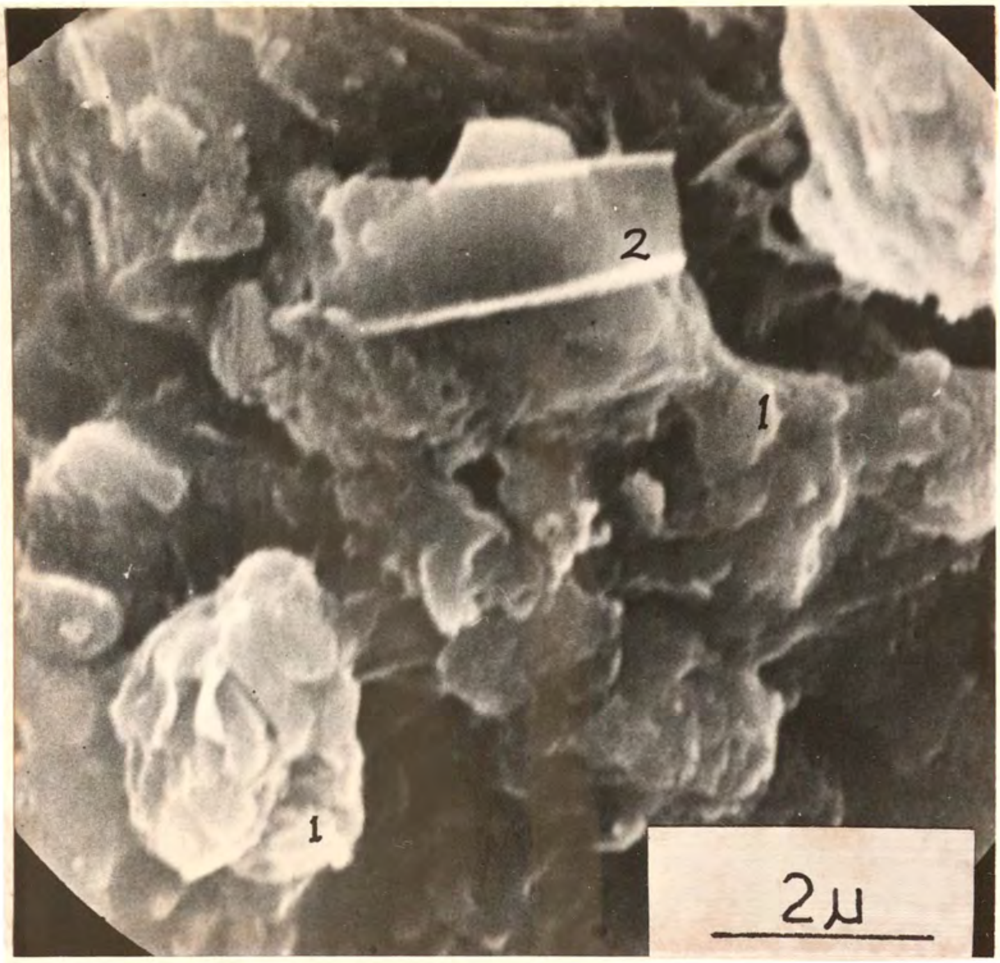
Plate 4.7. LANDSAT image (30 October, 1975; Band 4) of the inner Hamilton and North Taranaki Shelf areas. In the north suspended sediment emerging from the Aotea and Kawhia Harbours shows drift to the north. The dense ellipse-shaped plume emerging from the Mokau River extends southwestward onto the shelf suggesting the presence of a southerly surface current that is opposite in direction to the major Westland Current and is perhaps a backward gyre of that current (cf. Fig. 4.2.). More detailed analysis (Male, 1977) has shown that this current is not wind generated, and that the less dense southwesterly directed sediment plume immediately to the north is probably a tide-offset relict Mokau River plume. Scale 1 cm : 13 km.

Plate 4.8. LANDSAT image (15 February, 1976; Band 4) showing suspended sediment emerging from the Aotea and Kawhia Harbours and from the rivers bordering the North Taranaki Shelf. Note how sediment emerging from the rivers shows a prominent southerly drift, while sediment emerging from the Kawhia and Aotea Harbours shows a northerly drift (cf. Plate 4.7 and Fig. 4.2.). Scale 1 cm : 13 km.

Plate 4.9. Scanning electron micrograph of a Hamilton Shelf seston sample (N377). Note the abundance of diatomaceous material (1) and the floccules of predominantly montmorillonitic clay material (2).

Plate 4.10. Higher resolution scanning electron micrograph of a Hamilton Shelf seston sample (N377). The clay material is dominantly montmorillonite (1) (cf. Plates 3.9 and 3.10). Note the glass shard (2).


2
1
2
1
20μ


2
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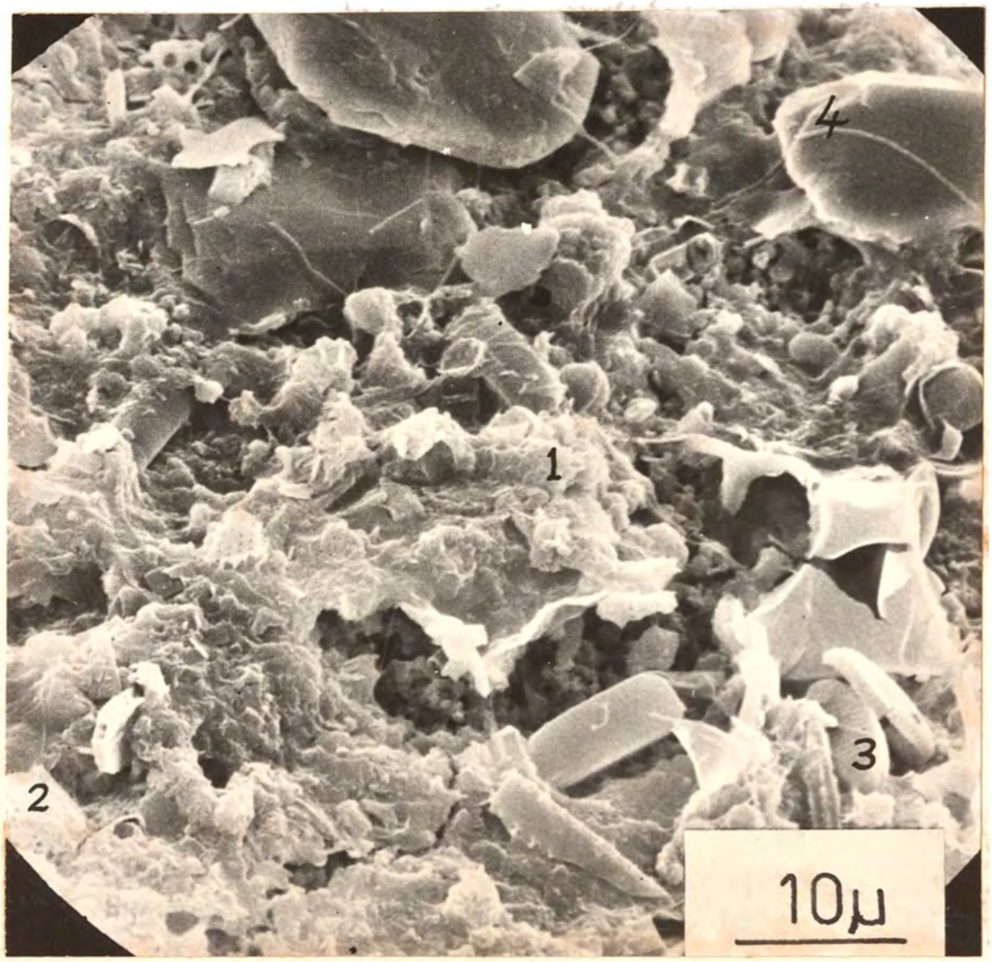
4
1
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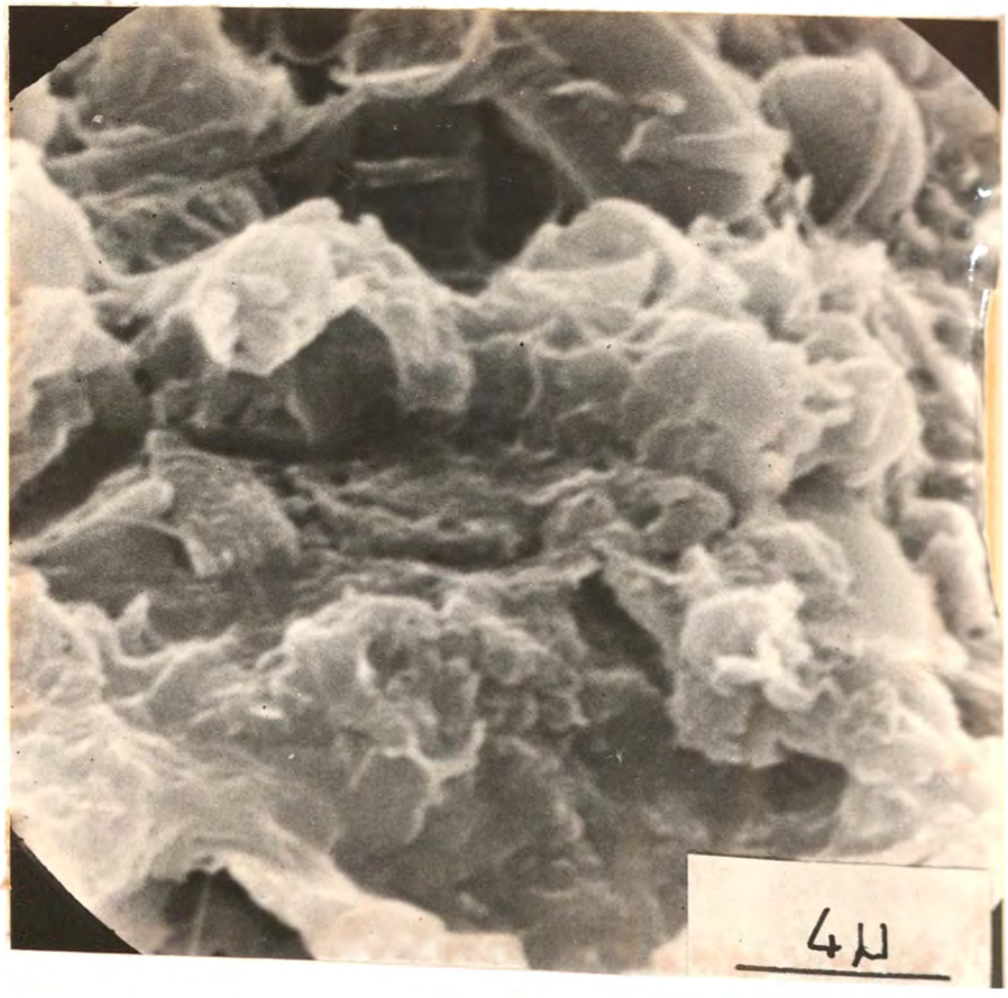
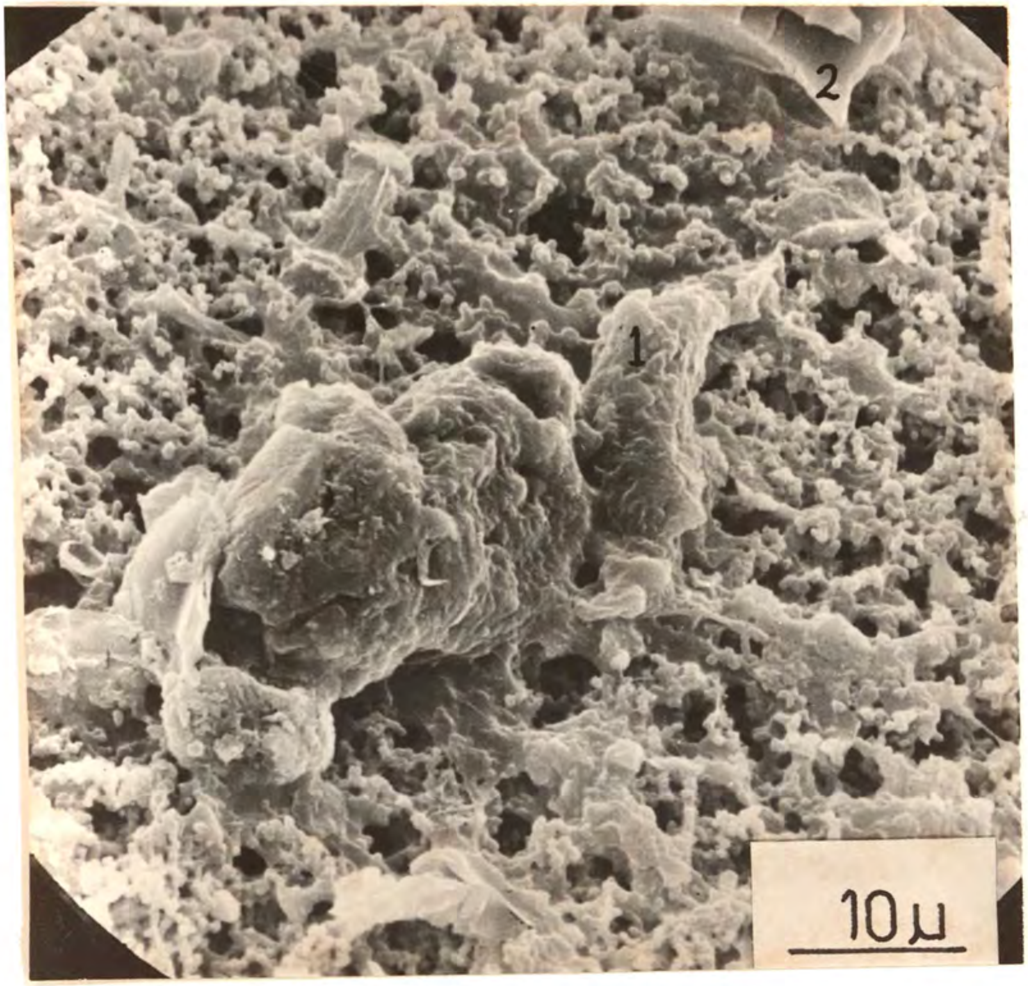

4µ

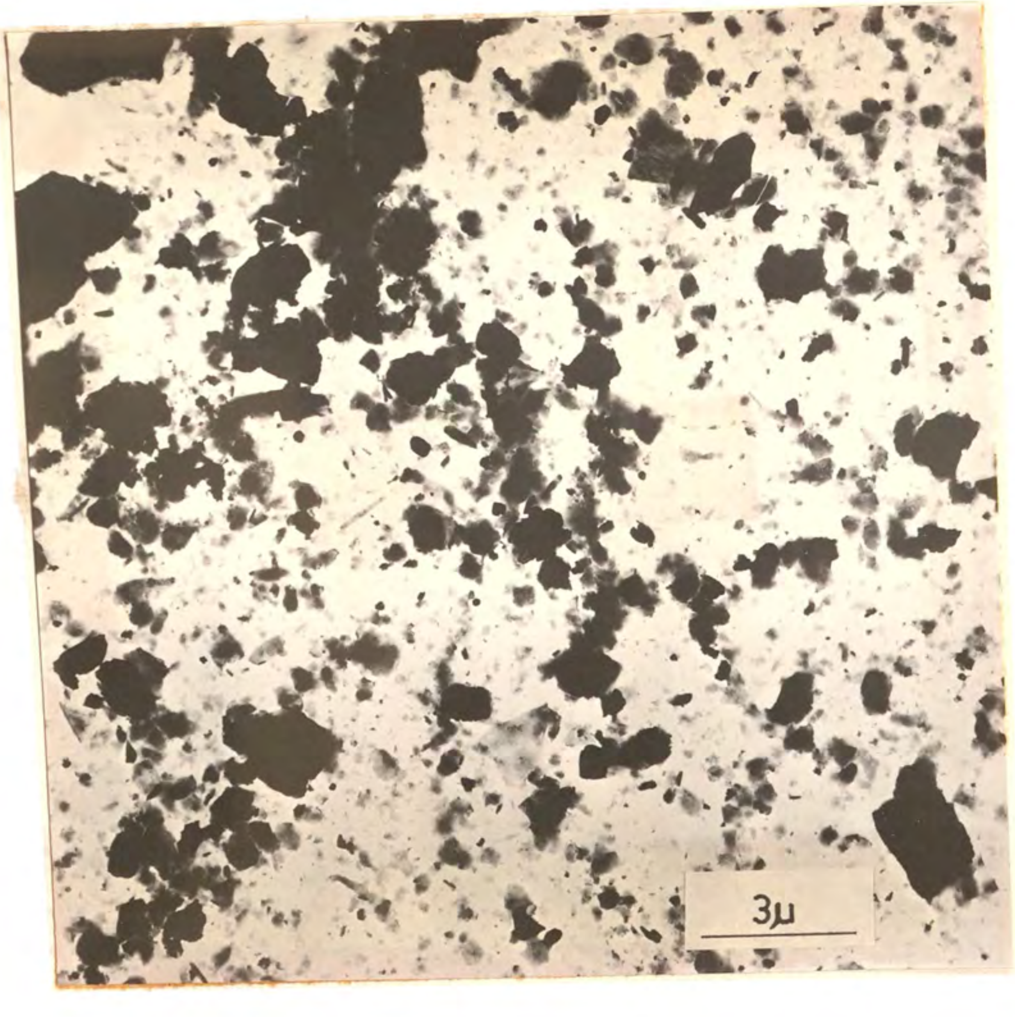
Plate 4.11. Scanning electron micrograph of a Hamilton Shelf seston sample (N378). The sample contains abundant clay material (1), most of which is montmorillonite, diatomaceous material (2), coccolith material (3), quartz and feldspar minerals (4).

Plate 4.12. Higher resolution scanning electron micrograph of a Hamilton Shelf seston sample (N378) showing abundant montmorillonite clay material.

Plate 4.13. Scanning electron micrograph of a Hamilton Shelf seston sample (N379) showing unidentified amorphous material (1) and a quartz grain (2). Note the typically low abundance of material in the sample. The millepore filter media forms the background.

Plate 4.14. Transmission electron micrograph of typical clay fraction ($< 2\mu$) material from a North Taranaki Shelf surficial sediment sample (C275). Dark grains represent clusters of chloritic and illitic material. Note the relative abundance of angular illitic and chloritic material to fine-grained montmorillonite particles.


2
1
10µ


3µ

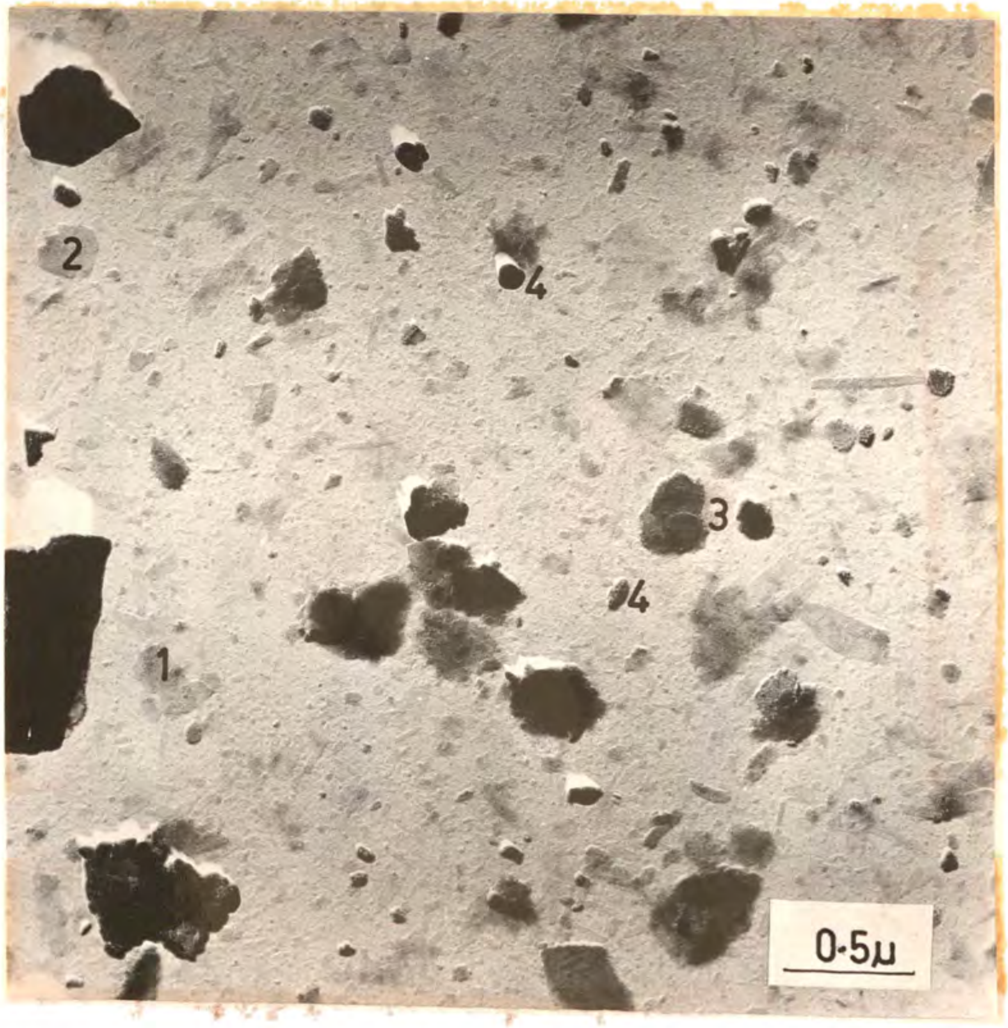
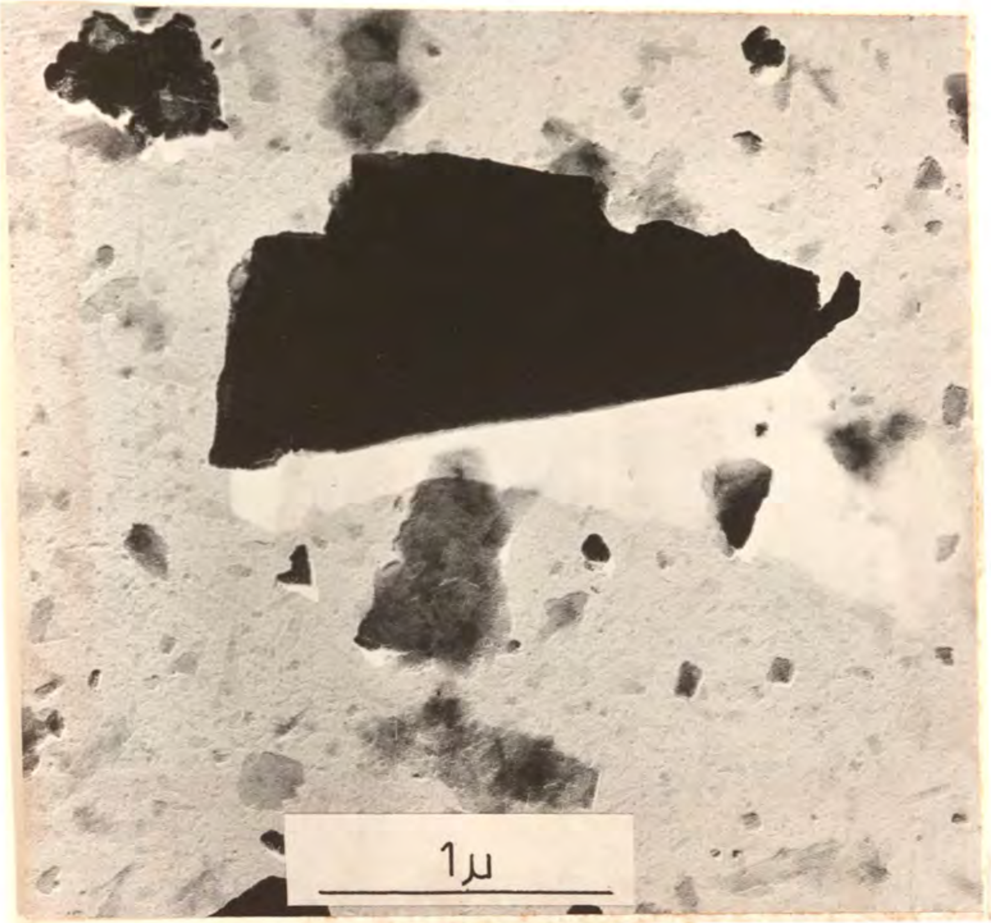

2
4
3
4
1
0·5µ

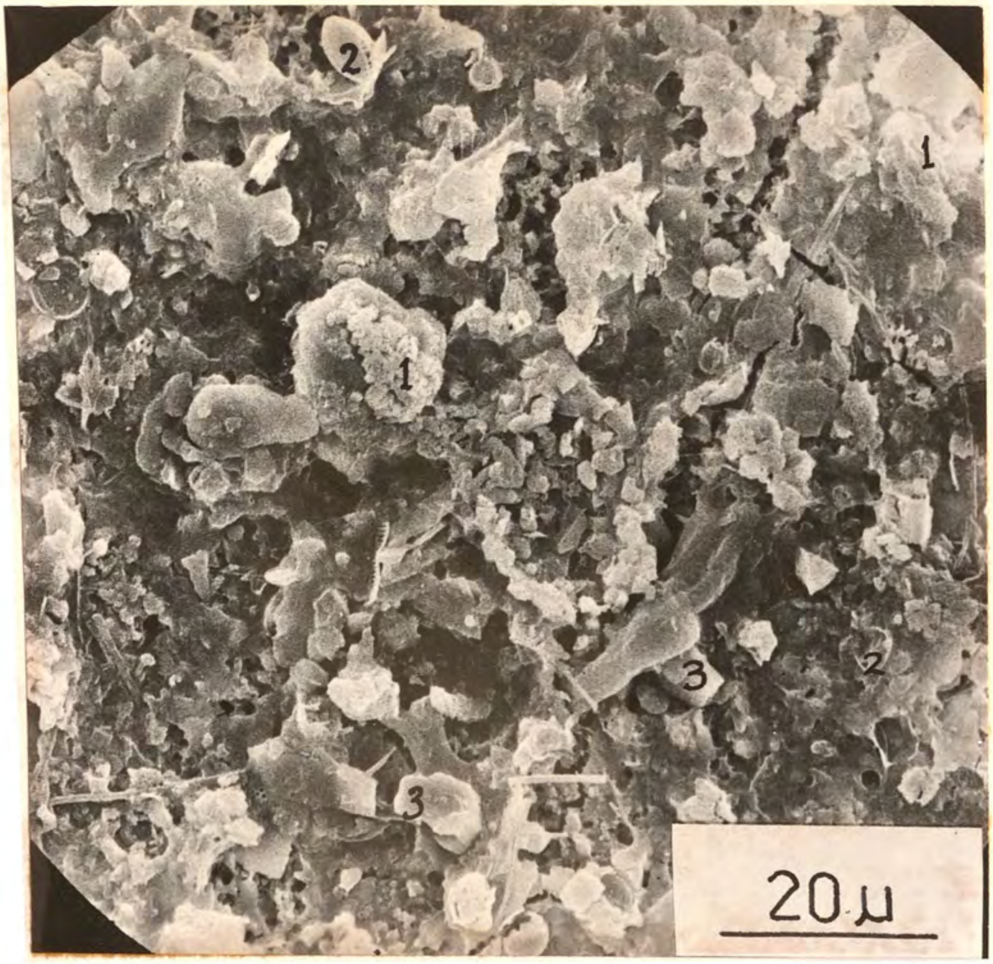
Plate 4.15. Grass covered slopes in foreground show the typical geomorphologic surface features produced by erosion in the montmorillonite-rich Taumatamaire mudstones. Steep bluffs in the background are formed from more stable Mokau and Mohakatino Group sediments. Photograph taken in the Mahoenui area looking southeast from N91/388519 towards the Aria Fault Scarp.

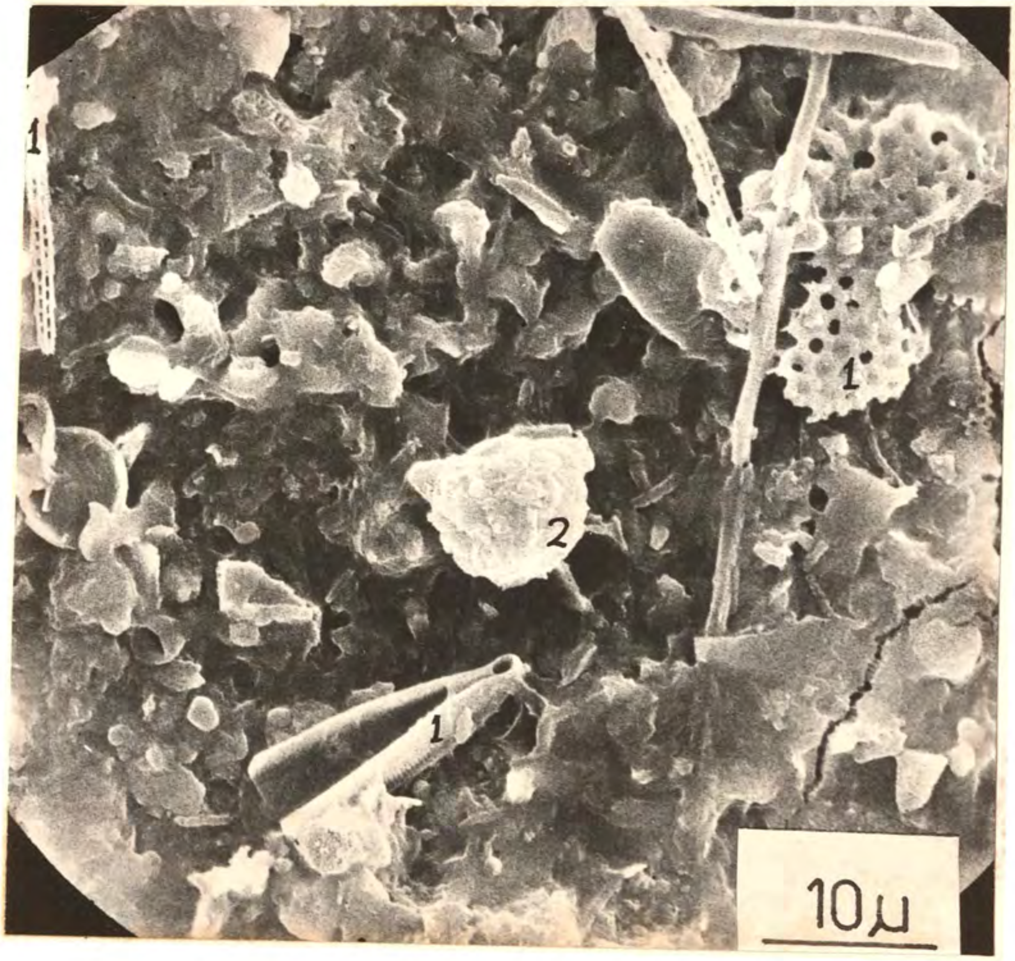
Plate 4.16. Transmission electron micrograph of clay fraction ($< 2\mu$) material from a Central Taranaki Shelf surficial sediment sample (B795) showing fine-grained montmorillonite flakes (1), coarser illitic (2) and chloritic (3) material and rounded elongate particles (4) of amorphous siliceous material. Note the abundance of lath-shaped (neoformed?) montmorillonite flakes.

Plate 4.17. TEM of clay fraction ($< 2\mu$) material from a Central Taranaki Shelf surficial sediment sample (C363) showing coarse illitic and chloritic material (1), fine-grained montmorillonite flakes (2) and rounded and elongate particles (3) of amorphous siliceous material. Note the abundance of lath-shaped (neoformed?) montmorillonite flakes.

Plate 4.18. Scanning electron micrograph of a nearshore Central Taranaki Shelf sample (N382). The sample shows abundant clay (1) and biogenic material (2) and quartz and feldspar grains (3).


1μ


2
1
1
3
2
3
20μ


1
1
2
1
10µ

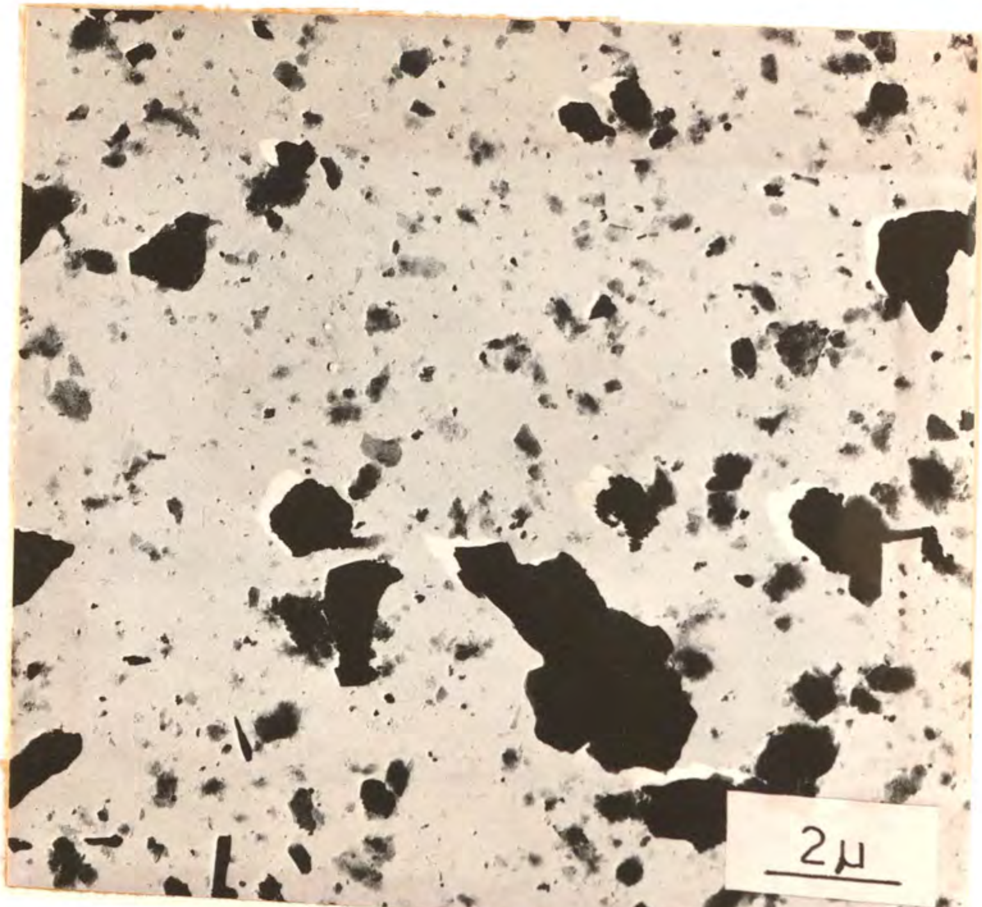

2µ

Plate 4.19. Scanning electron micrograph of a nearshore Central Taranaki Shelf sample (N383). The sample shows diatomaceous (1) and clay material (2).

Plate 4.20. Transmission electron micrograph of clay fraction ($<2\mu$) material from a South Taranaki Shelf surficial sediment sample (B645) showing an abundance of illite (1) and chlorite (2) material. The fine-grained background material is montmorillonite (3).

Plate 4.21. LANDSAT image (30 October, 1975; Band 4) of Cook Strait and adjacent areas which shows that currents move sediment south along the west coast of the North Island and eastwards through Cook Strait, and eastwards around the northern coastline of the Marlborough Sounds and through Cook Strait (cf. Fig. 4.2). Sediment from the large northern South Island rivers is initially swept northwards but then is swept eastwards by the D'Urville Current as Cook Strait is approached. Scale 1 cm : 17 km.

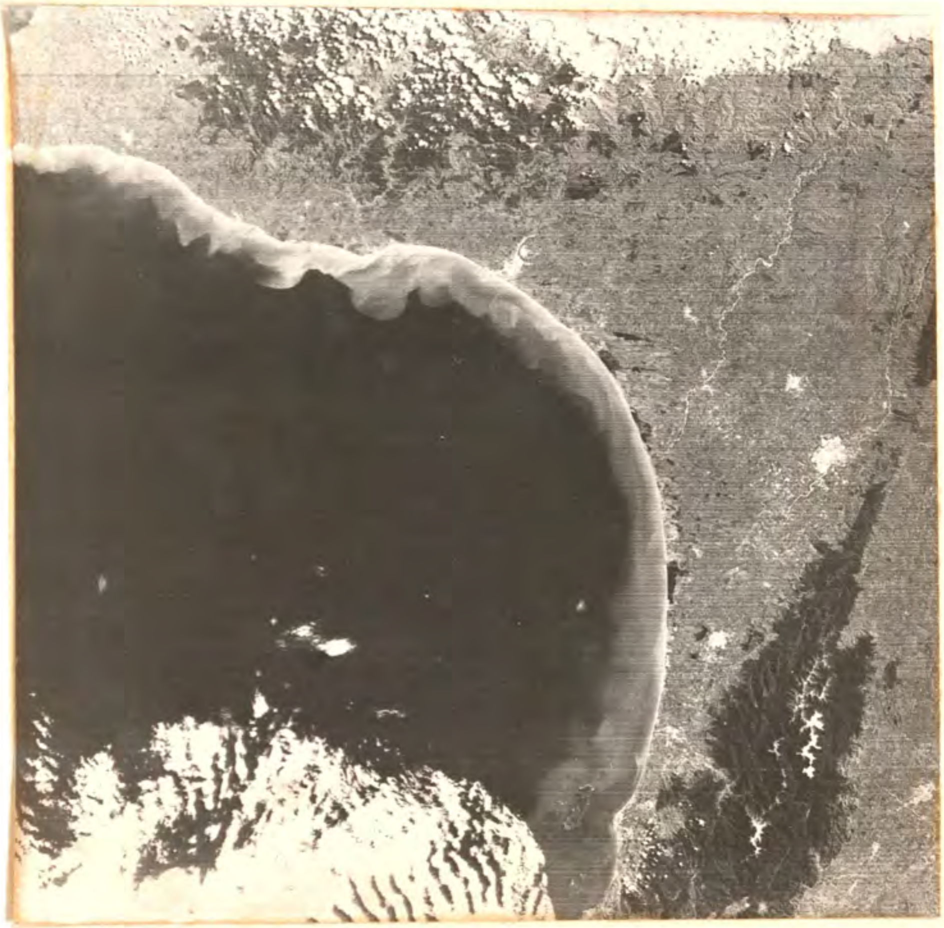
Plate 4.22. LANDSAT image (30 October, 1975; Band 4) showing suspended sediment emerging from the rivers and streams of the northwestern South Island (13 October, 1975). The Cape Farewell sandspit forms the northern boundary to Golden Bay in the bottom left, the northern tip of the Marlborough Sounds is the cloud covered area at the bottom right and Mt Egmont can be seen through a cloud gap to the right of top centre. (cf. Fig. 4.2). Scale 1 cm : 14 km.

Plate 4.23. LANDSAT image (17 February, 1976; Band 4) showing suspended sediment drifting northeast from the rivers and streams of the northwestern South Island. Scale 1 cm : 13 km.

Plate 4.24. Geomorphic surface features produced by erosion in the Tangahoe Mudstone in hill country 1 km south of Otomoa (N138/772040; Locality 78). Note the grooved patterns produced on the flanks of the grass-covered mudstone slopes by mass movement.

Plate 4.25. LANDSAT image (17 November, 1975; Band 4) showing suspended sediment emerging from the rivers and streams draining into the South Taranaki Bight. In central and western areas sediment plumes extend some 10 km directly onto the shelf but south from the Wanganui River sediment shows a marked southerly drift, a feature shown in other LANDSAT images of this part of coastline (cf. Plate 4.26). Scale 1 cm : 18 km.

Plate 4.26. LANDSAT image (23 December, 1975; Band 4) showing suspended sediment emerging from the rivers and streams draining into the South Taranaki Bight. The sediment plumes extend about 10 km directly onto the shelf. Scale 1 cm : 16 km.



The following 11 pages were tucked into the back pocket of this thesis along with a copy of the following journal article:

Nelson, C. S., & Hume, T. M. (1977). Relative intensity of tectonic events revealed by the tertiary sedimentary record in the North Wanganui Basin and adjacent areas, New Zealand. New Zealand Journal of Geology and Geophysics, 20(2), 369–392. https://
doi.org/10.1080/00288306.1977.10420714

STRATIGRAPHY						GENERALISED LITHOLOGY	Plate Reference
Period	Epoch	Series	N.Z.Stage	Group	Formation		
Quaternary	Recent (Rc)					Volcanic ash and soils.	-
	Pleistocene	Hawera (H)				Silt, sand, gravel and volcanogenic debris.	-
			Castlecliffian (Wo)			Siltstone, sand, conglomerate and shell beds.	-
		Wanganui	Nukumaruan (Wn)			Limestone, siltstone, sandstone, shell beds and lignite bands.	-
Tertiary	Pliocene		Wo - Wm		Tangahoe Mudstone (Tg)	Massive blue-grey mudstone with some sandstone and concretions.	1.23, 1.24
			Tk - Wo		Matemateaonga Sandstone (Mt)	Blue-grey mudstone, sandstone, conglomerate, shell beds and some concretions.	1.22
	Up.Miocene	Taranaki	Up. Tt		Urenui Siltstone (Ur)	Micaceous blue-grey siltstone with some concretions.	1.21
			Lr.-Mid. Tt	Mohakatino	Mt.Messenger Sandstone (Mt,M)	Massive and bedded sandstone and lesser mudstone, slightly tuffaceous, with some coal beds and slump structures.	1.19, 1.20
			Lr. Tt		Taueriki Mudstone (Tw)	Blue-grey mudstone and sandstone tuffaceous in part.	1.18
	Mid.Miocene	Southland	Sw		Ferry Sandstone (Fy)	Non-tuffaceous sandstone with some mudstone and slump structures.	1.16, 1.17
			Sw - Lr.Tt		Omceo Formation (Om)	Alternating beds of micaceous sandstone and siltstone with minor volcanic detritus.	-
			Sc - Sw		Pukupuru volcanic Sandstone (Pu)	Tuffaceous sandstone and mudstone interbedded in part.	1.14, 1.15 1.16
	Lr.Miocene	Pareora	Mid.-Up. Pl		Mangarara Sandstone (Mg)	Tuffaceous calcareous sandstone with some limestone.	1.14
			Mid. Pl	Mokau	Upper Mokau Sandstone (U.Mo)	Blue-grey argillaceous fine sandstone with some conglomerates and concretions.	1.14
			Mid. Pl		Maryville Coal Measures (My)	Upper and lower coal zones and carbonaceous shale separated by carbonaceous sandstone.	-
			Lr.-Mid.Pl		Lower Mokau Sandstone (L.Mo)	Massive, ferruginous, argillaceous sandstone, carbonaceous or thin coal seams.	1.11, 1.12 1.13
			Po - Pl	Mahoenui	Taumarunui Formation (Tmr)	Flysh-type graded beds of alternating sandstone and mudstone.	1.10
			Up.Lw - Po		Taumatamaire Formation (Tmt)	Massive blue-grey mudstone with some sandstone and limestone.	1.9
	Oligocene	Landon	Lw	Te Kuiti	Otorohanga Limestone (Ot)	Crystalline flaggy limestone.	1.7, 1.8
			Ld - Lw		Waitomo Sandstone (Wt)	Massive calcareous sandstone.	1.7
			Ld - Lw		Orahiri Limestone (Or)	Sandy limestone, flaggy in part.	1.6, 1.7
			Ld		Mangapohi Sandstone (Mp)	Calcareous, glauconitic sandstone.	-
			Mid.Lwh-Ld		Te Akatea Siltstone (T.A)	Calcareous mudstone.	-
			Mid.Lwh-Ld		Aotea Sandstone (Ao)	Calcareous sandstone with some mudstone and limestone.	1.3, 1.5 1.7
			Lr.-Mid.Lwh		Whaingaroa Siltstone (Wh)	Massive blue-grey, calcareous siltstone with occasional limestone.	1.3, 1.4
			Lwh		Glen Massey Sandstone (G.M)	Calcareous mudstone, sandstone and limestone.	-
	Up.Eocene	Arnold	Ar - Lwh		Mangakotuku Siltstone (Mk)	Carbonaceous, purple-brown, non-calcareous siltstone with some concretions.	-
			Ar - Lr.Lwh		Waikato Coal Measures (Wk)	Carbonaceous mudstone, coal beds and some sandstone and conglomerate.	1.2
Jurassic		Oteke Kawhia, Herangi				Slightly indurated mudstone, sandstone and conglomerate with some tuff.	-
Triassic		Balfour				Indurated mudstone and sandstone with some tuff.	1.1

Table 1.1. Summary of the stratigraphy and lithology of major sedimentary units (note abbreviations in brackets) in the study area (after Kear and Schofield, 1959; Lensen <u>et al</u>., 1959; Kear, 1960; Hay, 1967; Happy, 1971) Where sedimentary structures, colour and degree of induration of lithologies are specified, the property is generally diagnostic of the bulk of the stratigraphic unit.

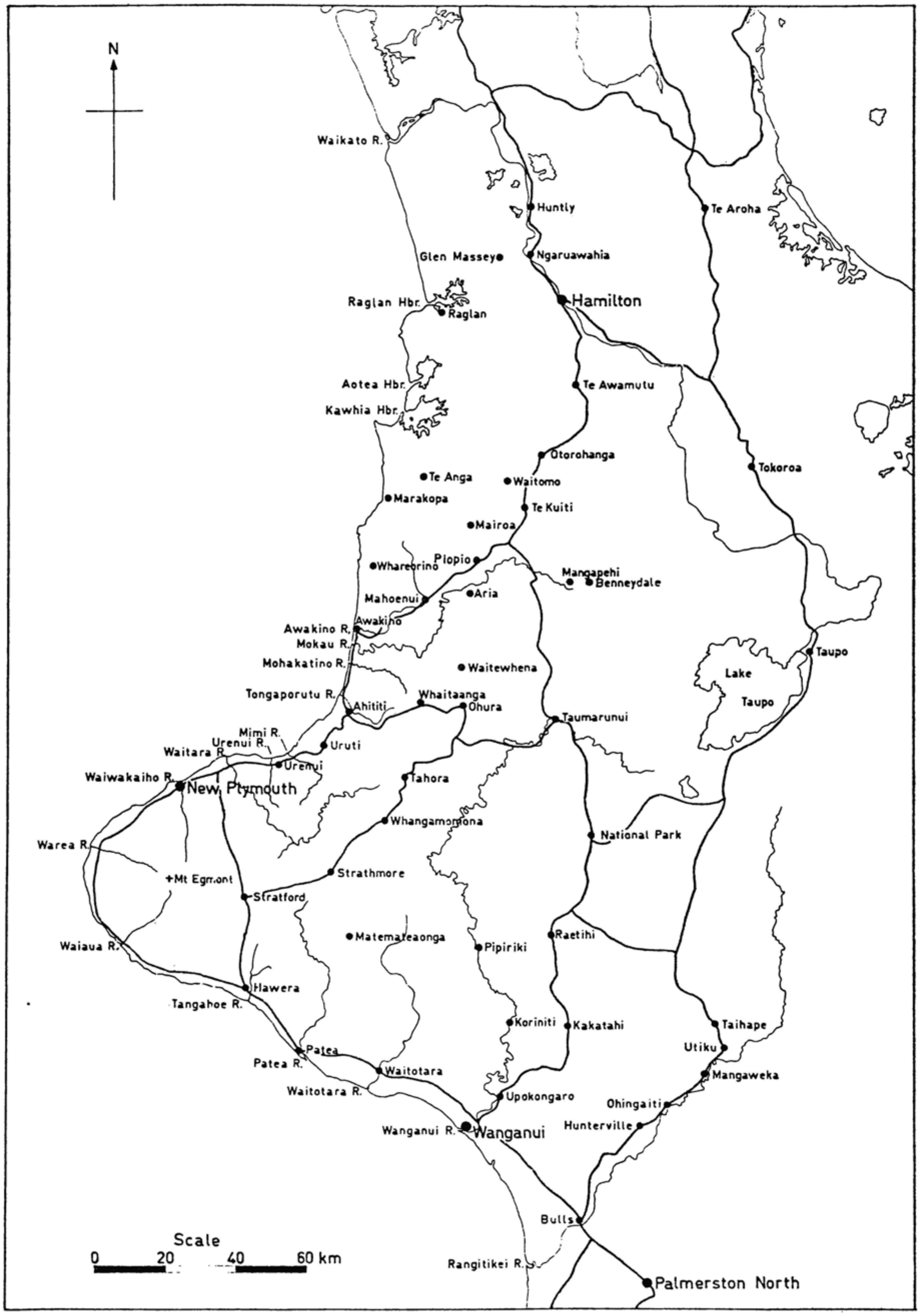


Fig. 1.2. Major place names, physiographic features and roads in the study area.

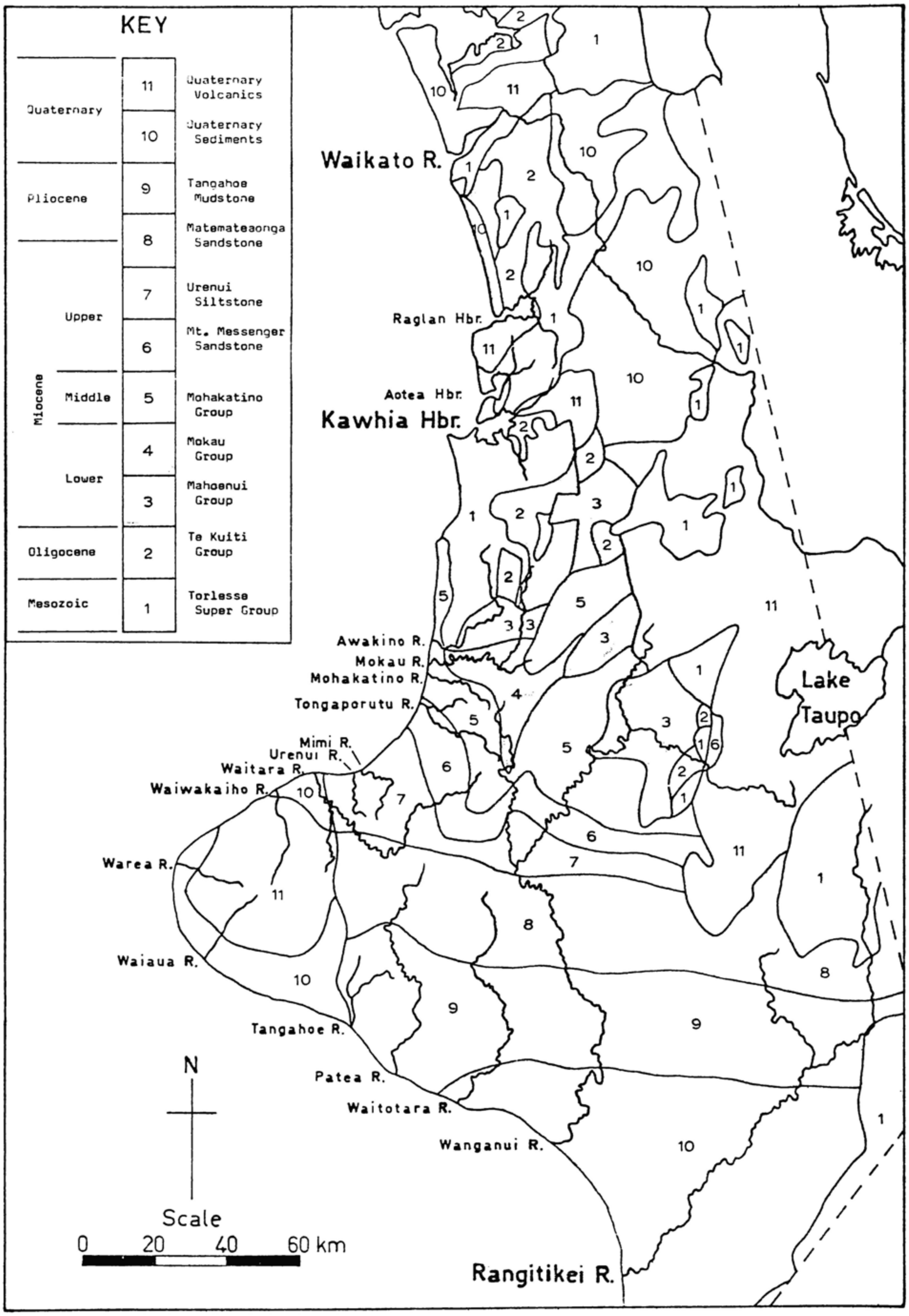


Fig. 1.3. Generalised geologic map of the onshore study area.

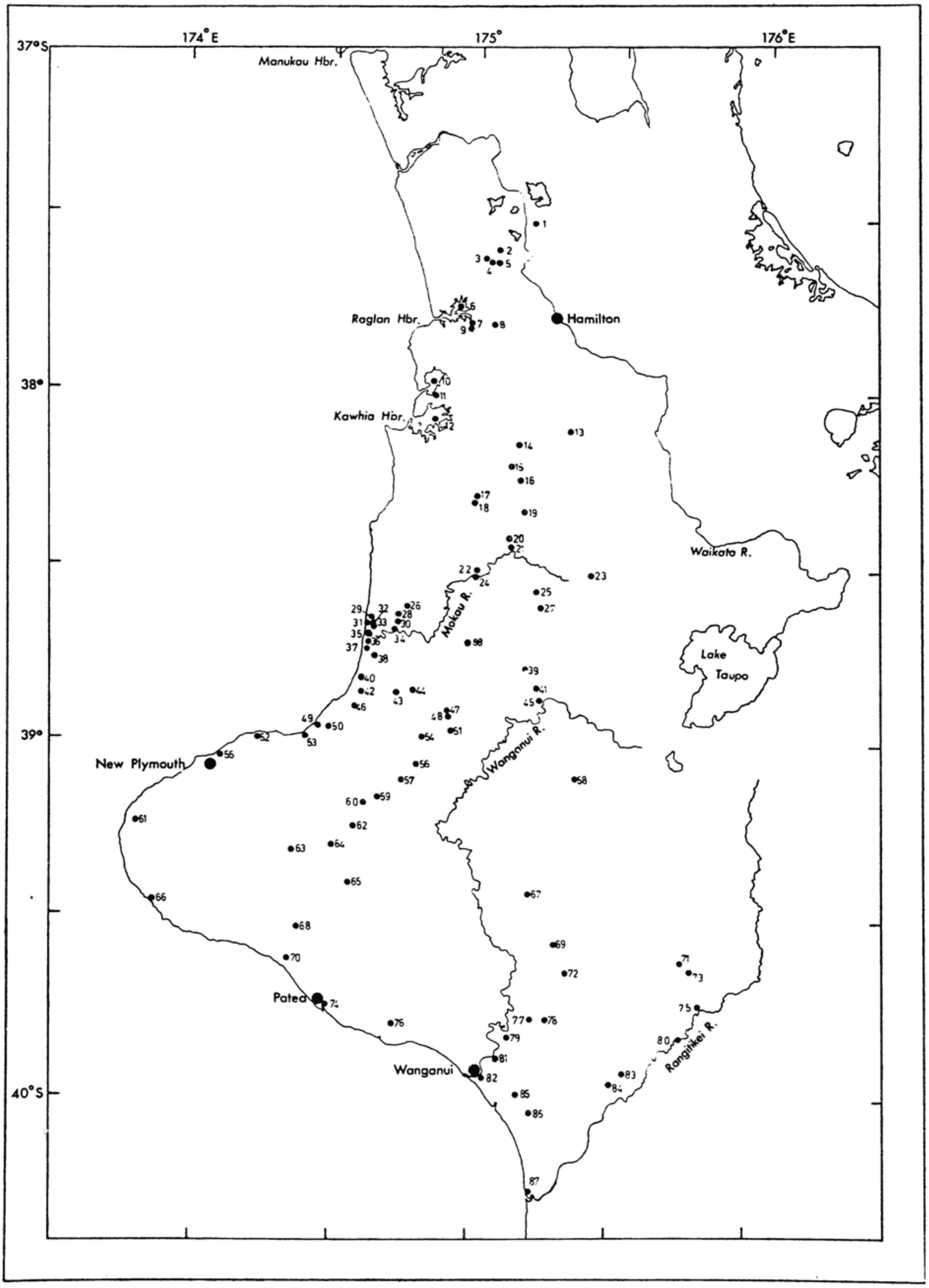


Fig. 2.1. Sample localities of onshore and harbour and river samples.

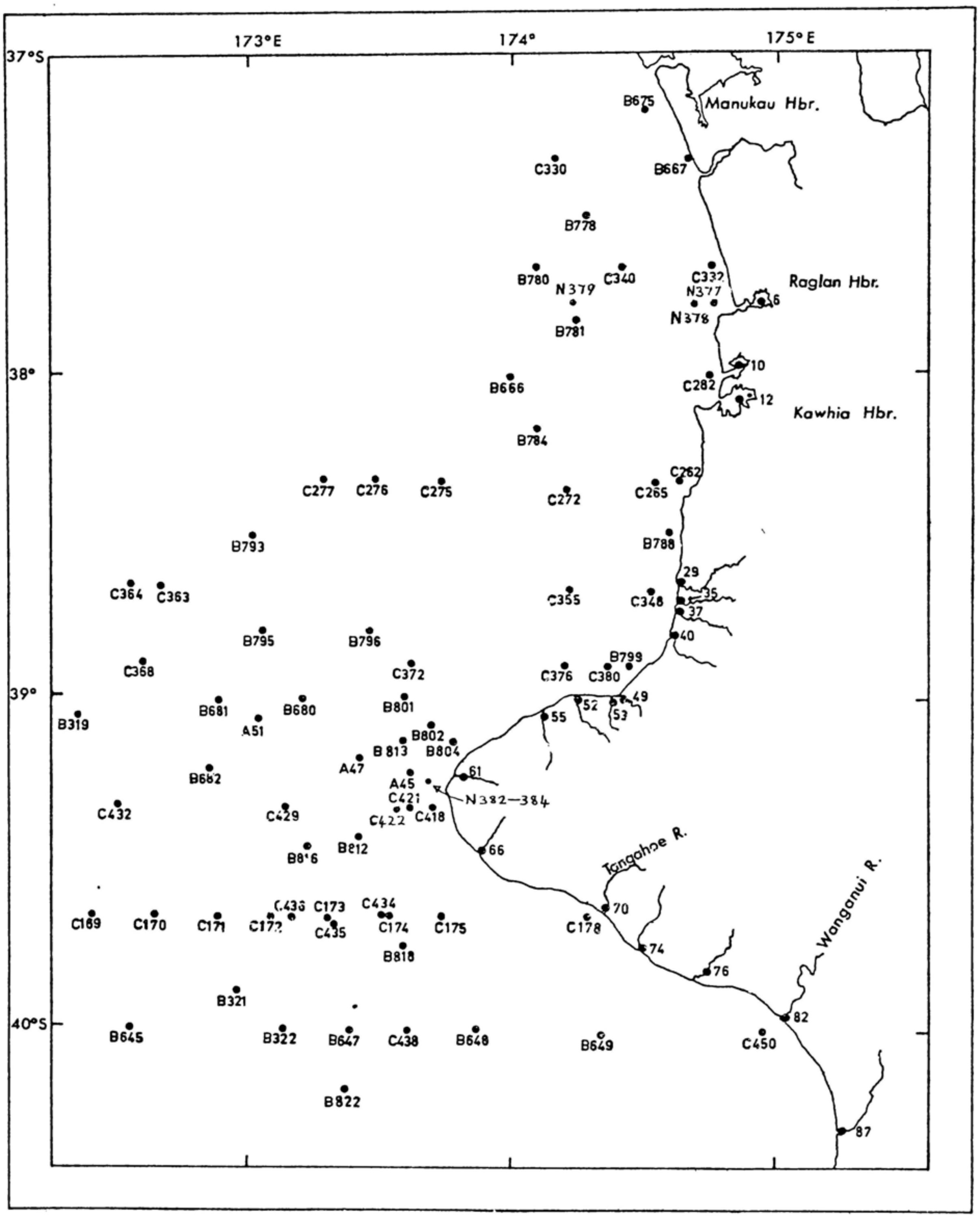


Fig. 2.2. Sample localities of western shelf and harbour and river samples.

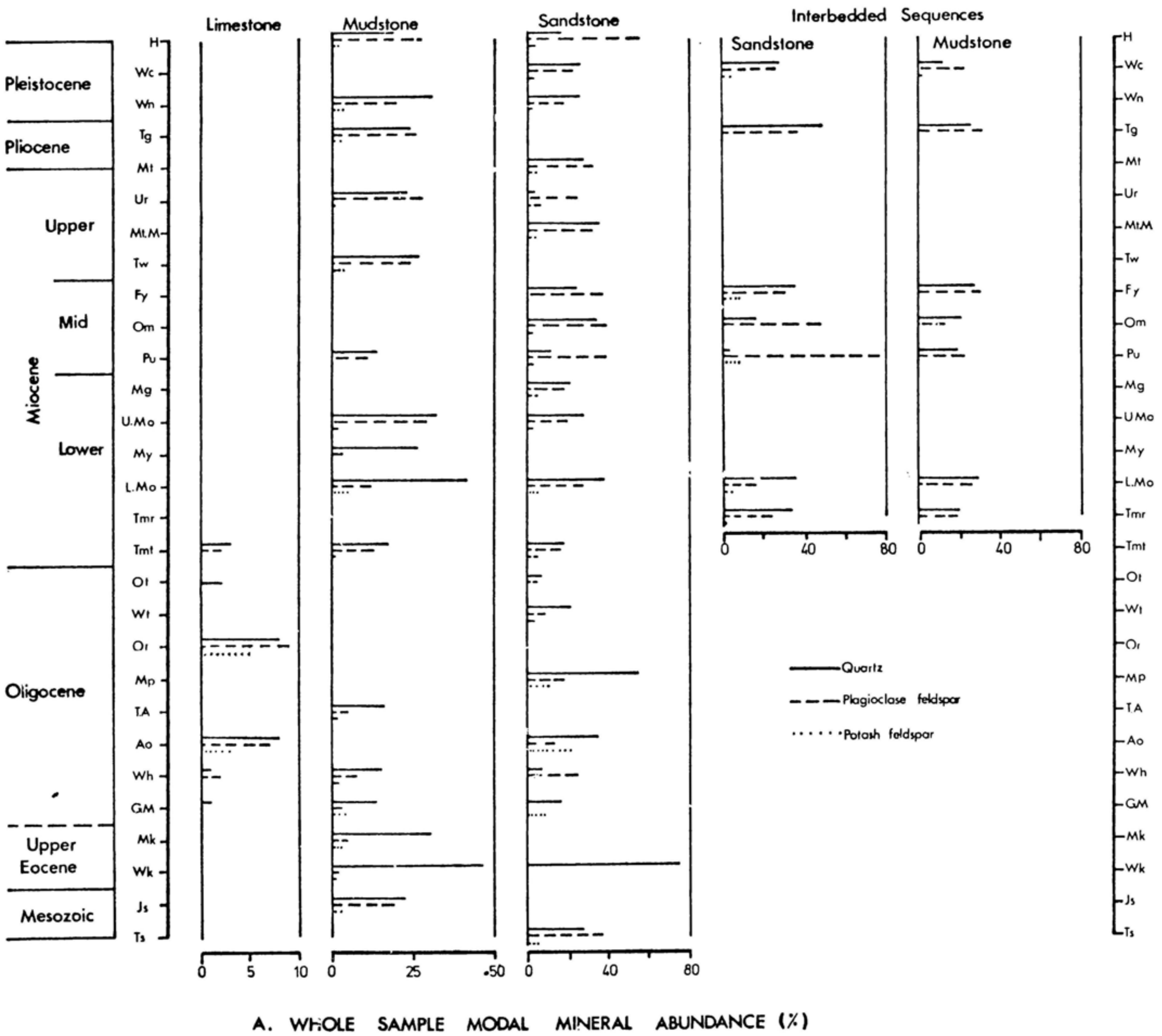


Fig. 3.4A. Bulk variations in average mineral compositions for the major sedimentary units in the onshore study area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1.

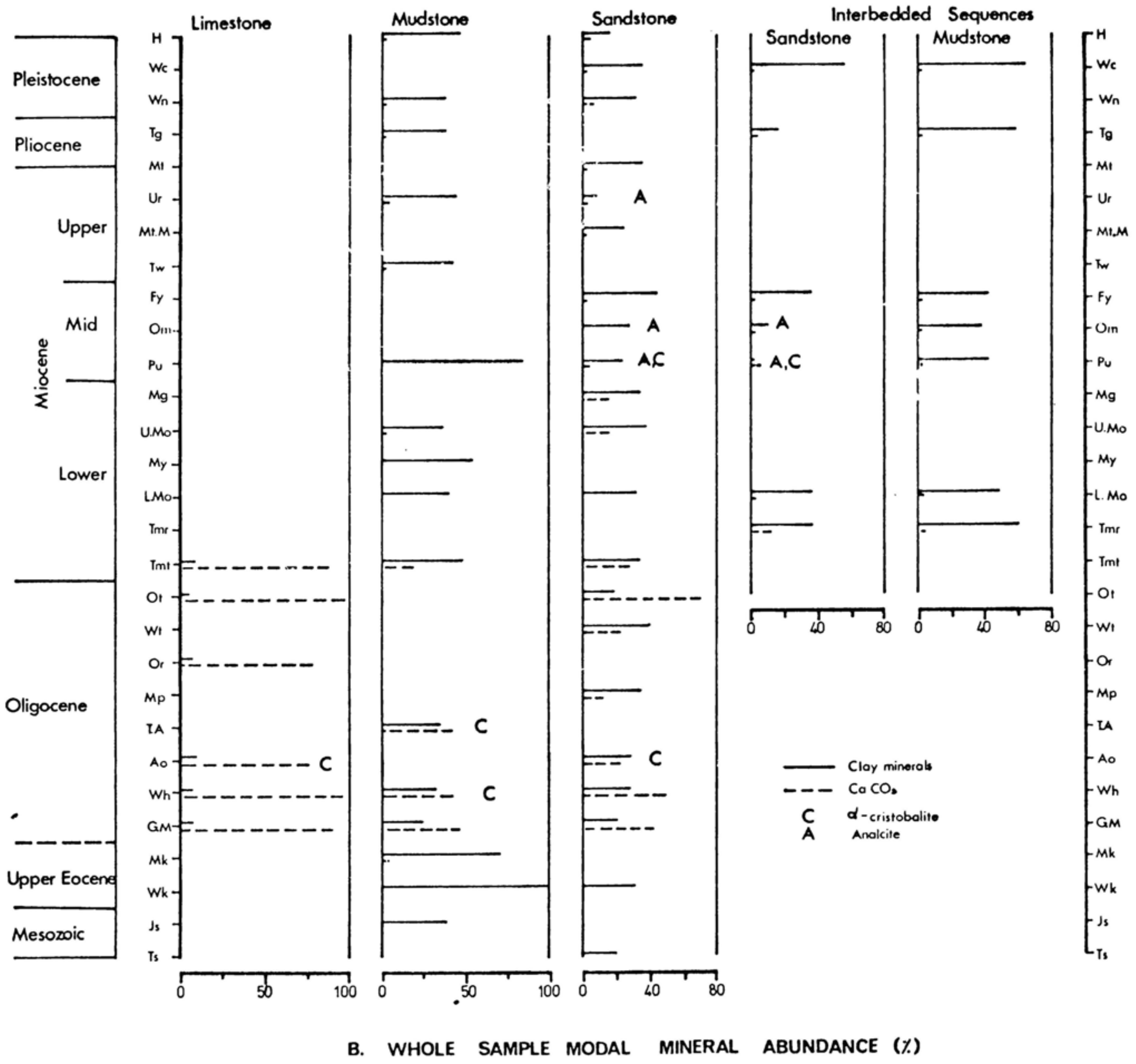


Fig. 3.4B. Bulk variations in average mineral compositions for the major sedimentary units in the onshore study area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1.

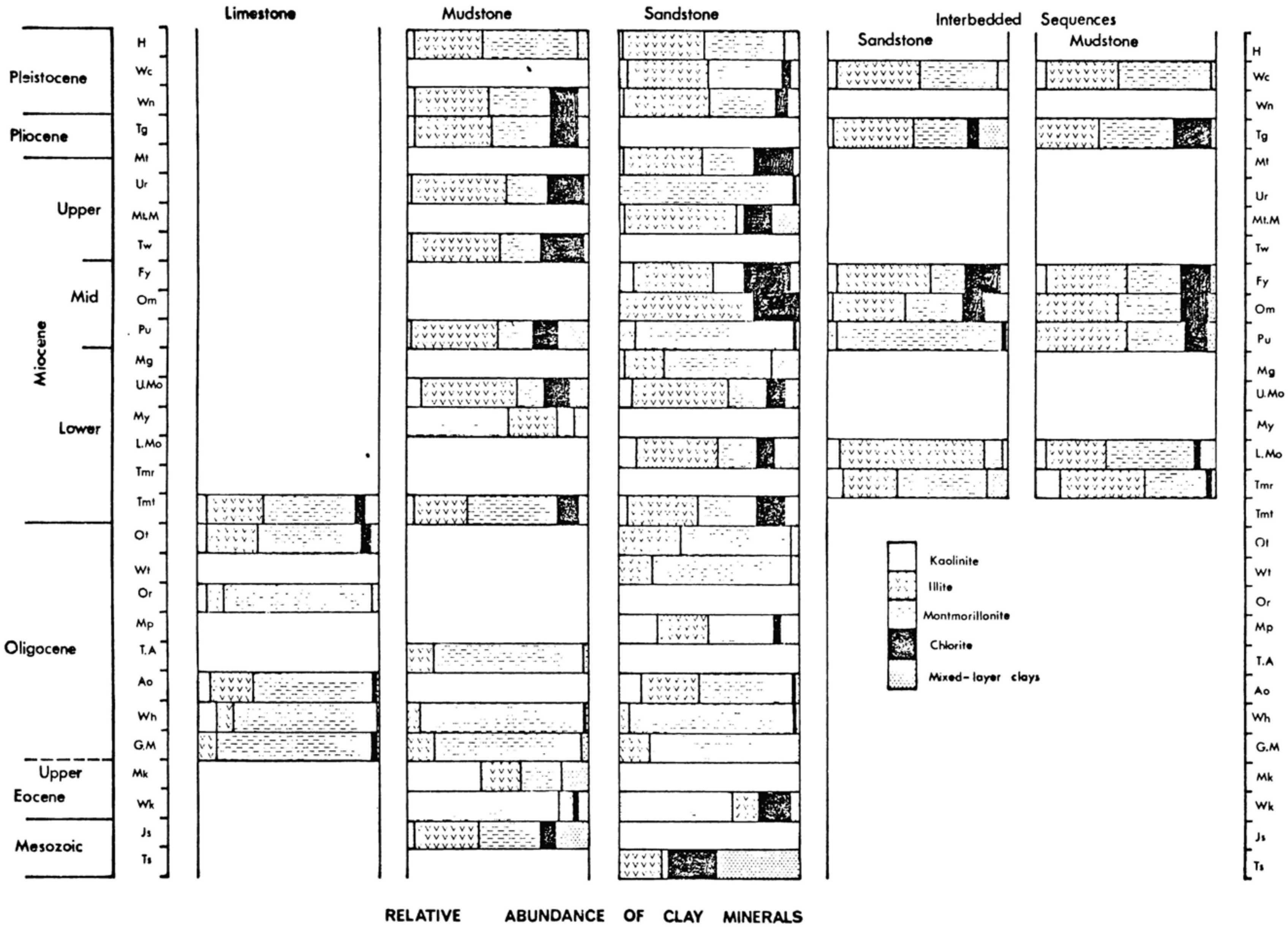


Fig. 3.5. Bulk variations in average clay mineral compositions for the major sedimentary units in the onshore study area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1.

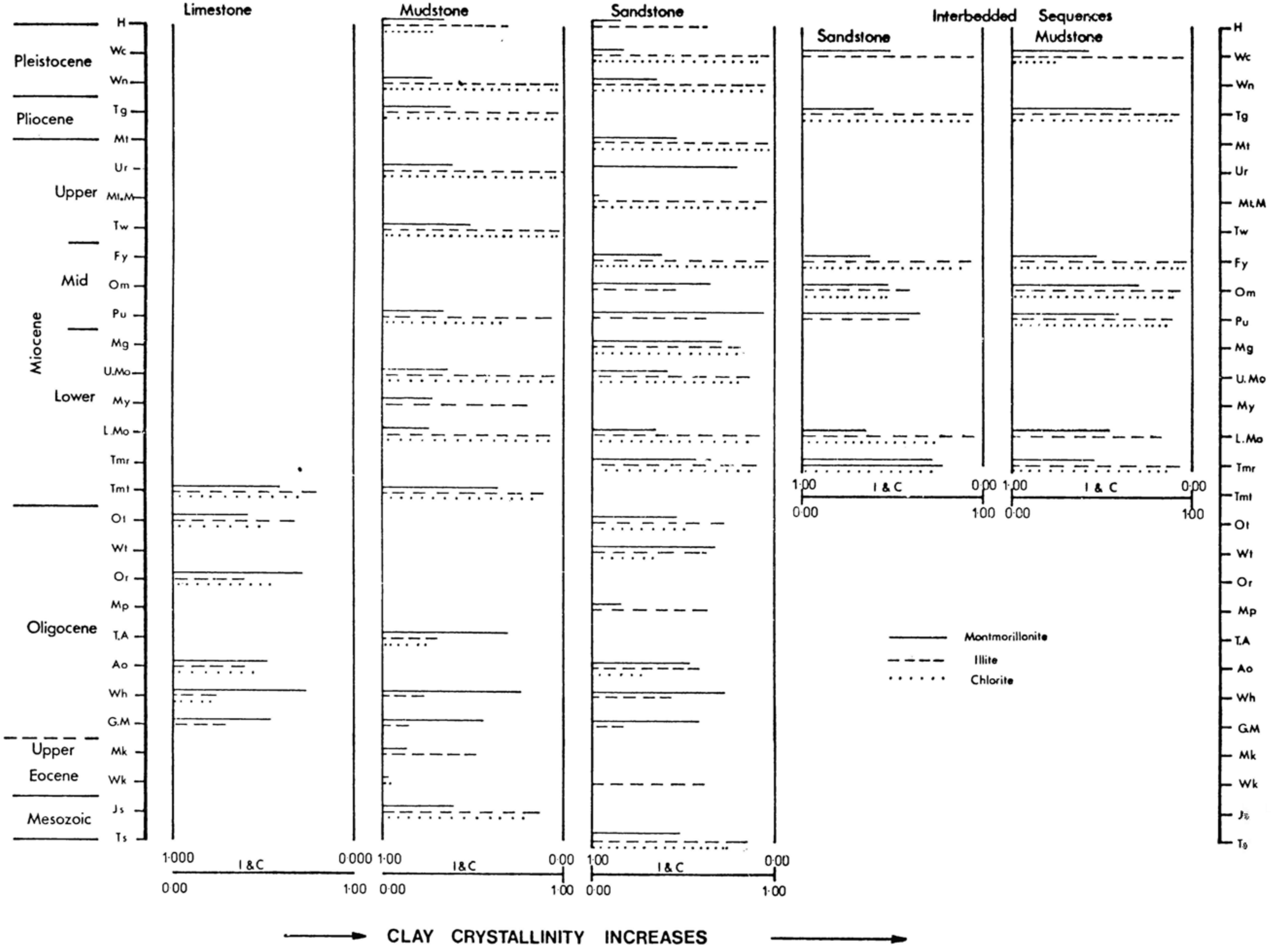


Fig. 3.6. Bulk variations in average clay mineral crystallinity for the major sedimentary units in the onshore study area as determined by XRD analysis. Abbreviations of formation names defined in Table 1.1.

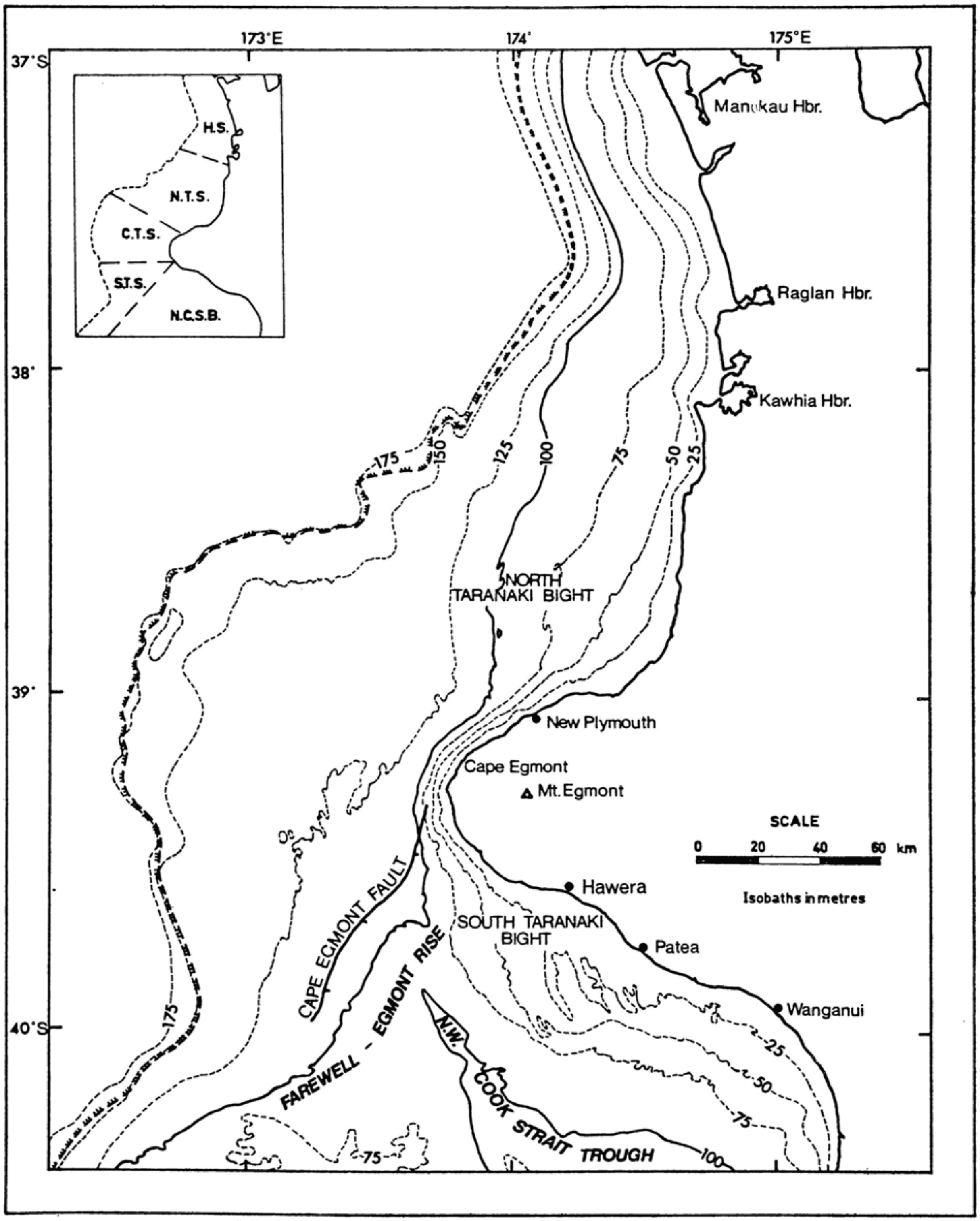


Fig. 4.1. Bathymetry and main physiographic features of the western continental shelf off the central Western North Island. Inset map shows the division of the western shelf into the Hamilton Shelf (H.S.), North Taranaki Shelf (N.T.S.), Central Taranaki Shelf (C.T.S.), South Taranaki Shelf (S.T.S.) and North Cook Strait Basin (N.C.S.B.). Edge of continental shelf marked by ╨╨╨╨╨ . (After McDougall and Brodie, 1967; Lewis and Eade, 1974).

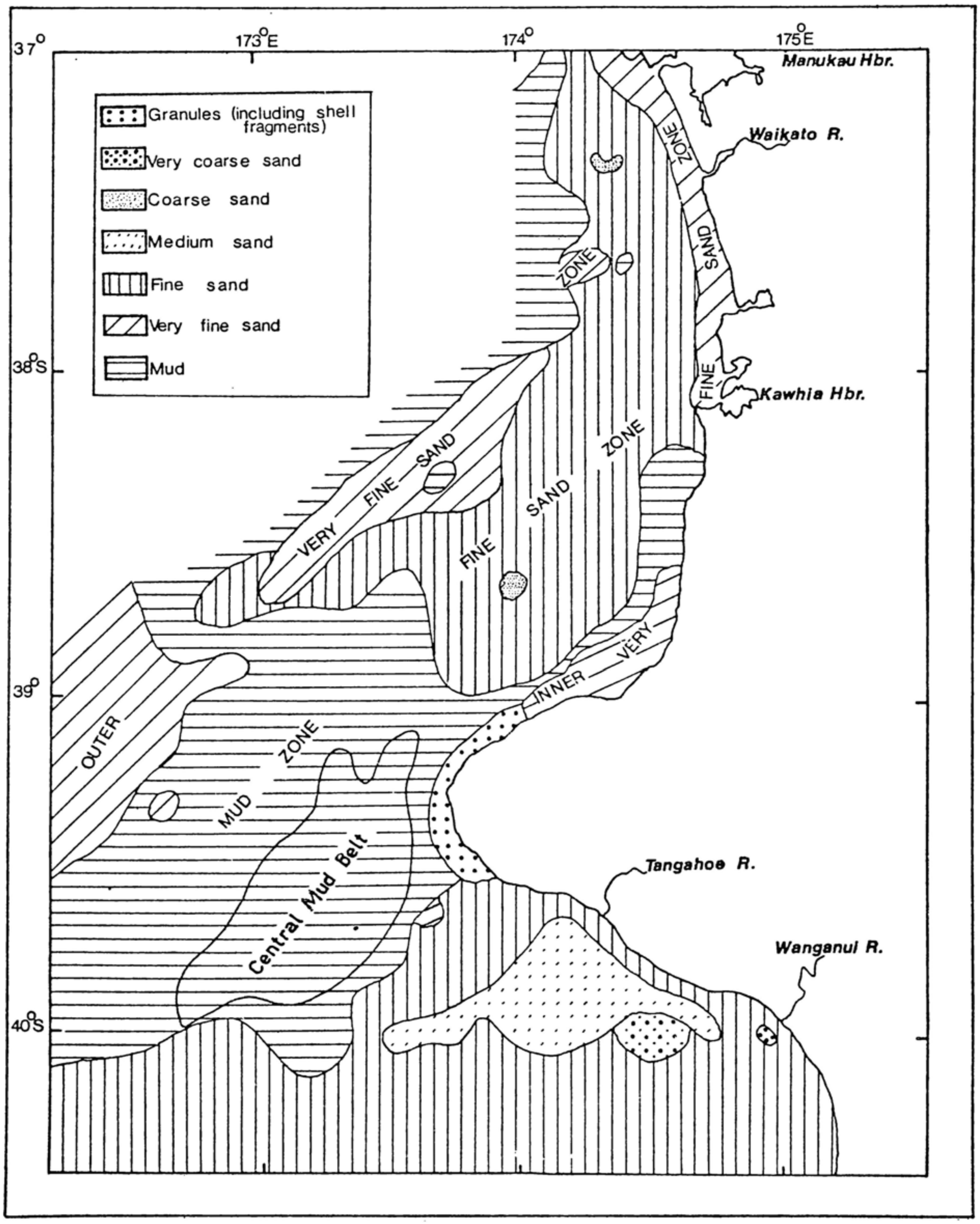


Fig. 4.3. Distribution of modal sizes in the sediments of the western continental shelf (after McDougall and Brodie, 1967; Lewis and Eade, 1974).