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Environmental and sea-level changes on Banks Peninsula (Canterbury, New Zealand) through three glaciation–interglaciation cycles

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Abstract

A greater than 200 ka record of marine transgressions and regressions is recorded from a 75 m core from Banks Peninsula, Canterbury, New Zealand. This record comprises thick suites of muddy sediments attributed to back barrier, lake and lagoonal environments alternating with thin soil and loess complexes. These deposits have been dated using radiocarbon and thermoluminescence (TL) techniques supported by proxy data (diatoms, phytoliths, pollen and sedimentology). The aqueous deposits are attributed to three interglacials and an interstadial (Marine Isotope Stages 1, 5a, 5c, 6, and 7). The loesses and paleosols date to the intervening stadials (Isotope Stages 2, 5d (or 6?) and probably 8). On the basis of transgressive beach facies, back barrier swamps and barrier-blocked lake deposits, a partial sea-level curve including data from Isotope Stage 5 is presented. Our data indicate that Banks Peninsula has been tectonically stable over that period and we provide sea-level points that support the existing isotope curve during Stages 5 and 6. Detailed diatom records are limited to Isotope Stage 1 and the latter part of Stage 5. Diatom histories recorded from these stages are remarkably consistent. Both indicate a progressive floral change from marine types through freshwater colonising species to freshwater planktonic assemblages. These reflect parallel histories of coastal evolution during the two interglacials. In both cases, marine transgression in the early part of the isotope phase was followed by lagoon development implying that a gravel spit extended across the embayment from the west. This was succeeded by lake development when the lagoon was cut off by the juncture of the spit with Banks Peninsula. This lake deepened as the coast rotated into swash alignment and the spit was converted into a gravel barrier. The vegetation history of the site indicates that mixed podocarp broadleaf forests, similar to the pre-European flora of Banks Peninsula, occupied the region during Isotope Stages 1 and 7. This contrasts with the palynological interpretation of a marine record (DSDP Site 594) from off the Canterbury coast which suggested that Isotope Stage 7 was markedly cooler than the Holocene. During glacial periods, forest was eliminated and replaced by a tall shrubland of mixed montane and coastal affinities. © 1999 Elsevier Science B.V. All rights reserved.

Keywords: late Quaternary; coastal evolution; sea-level change; New Zealand; micropalaeontology; thermoluminescence dating

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1. Introduction

Despite an extensive history of Quaternary research in New Zealand, there is a paucity of long, late Quaternary records. This reflects both the highly dynamic New Zealand landscape, where most depositional environments are ephemeral on glacial–interglacial timeframes, and a tendency for research to focus on the glacial record in mountainous areas, where long sedimentary records are rare. In the absence of these records, knowledge of the overall late Quaternary history of terrestrial New Zealand is remarkably fragmentary. By contrast, considerable progress has been made in deciphering climate histories from the marine record in the New Zealand region (e.g. Nelson et al., 1985; Heusser and van de Geer, 1994). These records confirm the general correspondence of glacial–interglacial cycles in New Zealand with the Northern Hemisphere pattern but are of low temporal resolution. Direct transfer of the

marine record to terrestrial New Zealand is fraught with potential pitfalls and good terrestrial records are required to verify the marine story. Terrestrial records which can be correlated directly to the oceanic record are particularly valuable. Estuarine and lagoonal systems provide the possibility of such a correlation through direct evidence of marine transgressions and regressions. Barrier-blocked lagoonal systems are potentially very useful sites for sea-level investigations, as the fresh to brackish water environments behind the barrier are sea-level-controlled and largely depositional.

The Kaitorete ‘Spit’–Lake Ellesmere complex is a large barrier-blocked lagoon system that occurs on the southern flank of Banks Peninsula, Canterbury (Fig. 1). Water levels in Lake Ellesmere are currently managed to prevent inundation of agricultural land but in historical times the lake extended into the valleys around the southwestern flank of the peninsula. These valleys are steep sided and narrow mouthed

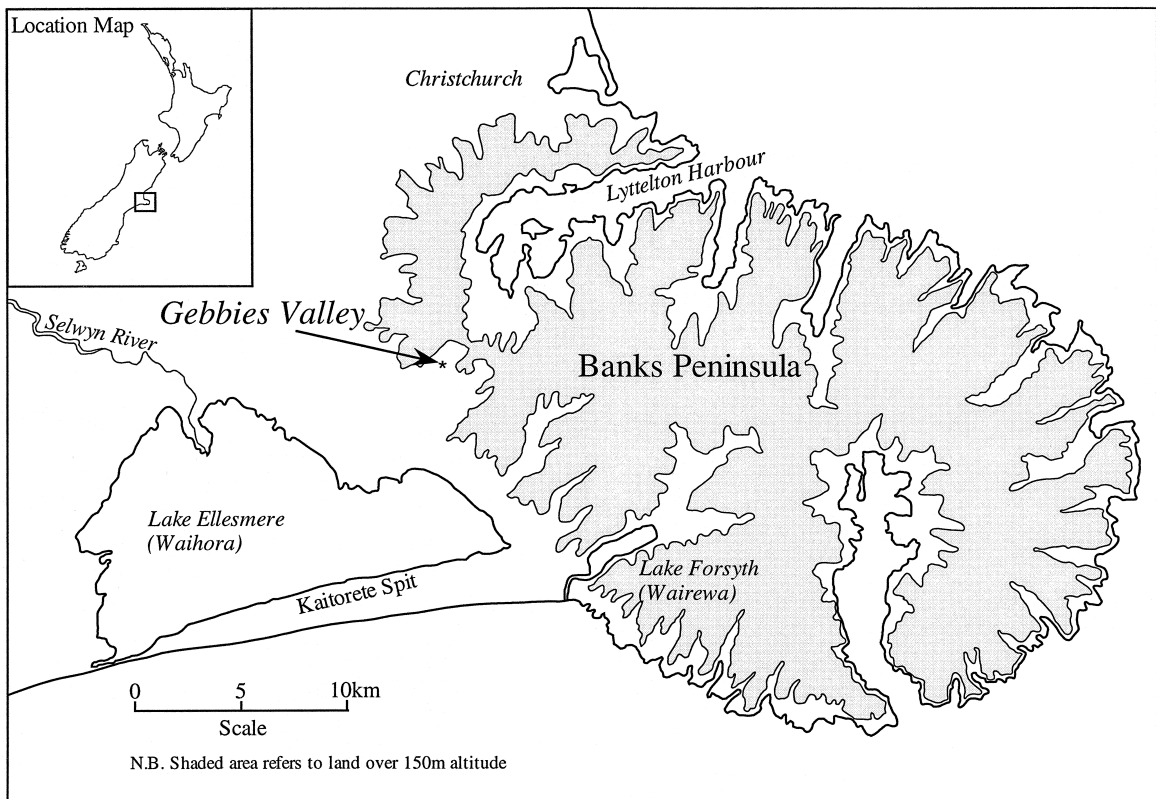


Fig. 1. Location map showing Banks Peninsula and the location of the Gebbies Valley drill hole. The drill site is marked with an asterisk.

and, even on open marine sections of the Banks Peninsula coast, provide loci for fine sediment deposition. Thick sediment fills exist in all these valleys. A drilling programme in Gebbies Valley along the southwest flank of peninsula, behind the Kaitorete 'Spit', recovered 75 m of largely fine-grained sediments. These sediments extend through the last >200 ka and are believed to represent three glacial–interglacial cycles, and record several marine transgressions and regressions. In this paper, we present an environmental history and the coastal evolution along the southwestern flank of Banks Peninsula for the last 200 ka.

1.1. Physiographic and geological setting

Banks Peninsula is the remnant of a complex of Miocene shield volcanos (Sewell, 1988). It is located along the eastern seaboard of the South Island, New Zealand, immediately south of the city of Christchurch (Fig. 1). The volcanic complex is largely basaltic and contains two large calderas, now occupied by Akaroa and Lyttelton Harbours. The complex has been heavily eroded and steep-sided valleys have been incised and/or exhumed along the outer flanks of the peninsula. These valleys probably first developed during the early history of the complex (Shelley, 1989), and the two calderas were formed before the deposition of the youngest lava flows 7–5.8 million years ago (Sewell et al., 1992).

Initially an offshore island, Banks Peninsula has been alternately attached to, or separated from, the mainland of the South Island of New Zealand, depending on interglacial–glacial variations in sea level, and on the extent to which the alluvial fans of the Canterbury Plains have been built out from the eastern front of the Southern Alps (Brown and Wilson, 1988). These fans are the source of material for the construction of barrier beaches, dunes and lagoons along their seaward edge which have linked the peninsula to the mainland.

On the southwest side of the peninsula, from Christchurch to Lake Forsyth, basement rock is 80–100 m below present sea level (Brown and Weeber, 1992, 1994), and the lower parts of the valleys are filled with sediment. The considerable depth of these valleys below present sea level may be related to incision during low glacial sea levels, or to subsidence

during the early history of Banks Peninsula. Wellman (1979) suggested that subsidence occurred at a rate of 0.2 mm/year, and about 600 m of sediment is recorded in the topographic low on the adjacent Canterbury Plains, now occupied by Lake Ellesmere (Talbot et al., 1986). However, Lawrie (1993) points to the presence of buried shore platforms at +5 to +8 m AMSL (above mean sea level), of probable Last Interglacial age, around the southwestern side of the peninsula and similar features have now been identified elsewhere on the peninsula (Bal, 1997). These imply local tectonic stability for at least the last 125,000 years.

1.2. General late Quaternary history

Two major types of Quaternary sediments are recorded on Banks Peninsula: loess and valley fills. Loess is widespread, characteristically forming a thin cover on upper valley sides and spurs, but increasing to several metres in depth on lower slopes. It is derived mainly from the Southern Alps greywackes, but the occurrence of sponge spicules in deposits on the eastern side of the peninsula indicates a subsidiary source on the exposed former seafloor to the south and east, during low sea-stands (Rae-side, 1964). Around the southwestern slopes of the peninsula loess overlies wave-cut platforms at +5 m AMSL, indicating a pre-Holocene age, and these are interpreted as Last Interglacial by Lawrie (1993). At least three loess layers separated by colluvium and palaeosols have been identified in this area and in Akaroa Harbour (Griffiths, 1973). A C-14 age of $17,450 \pm 2070$ yr BP was obtained for the uppermost layer in the Barry's Bay site in Akaroa Harbour (Griffiths, 1973), but Goh et al. (1977) demonstrated that this and other ages which they obtained from untreated material in the palaeosols gave minimum ages. More recent studies yielded ages of over 140 ka in the lower loess units (e.g. Berger et al., 1994a) and it is clear that many of the shore platforms around the southwest flank of Banks Peninsula are at least Last Interglacial in age, but could be considerably older.

The valley fills on Banks Peninsula are dominantly composed of muds and very fine sands (e.g. Dingwall, 1974; Soons et al., 1997). This study is the first detailed investigation of a valley fill complex.

2. Methods

2.1. Core and sample recovery

The core was located at the mouth of Gebbies Valley on a relict foredune system of historical Lake Ellesmere (see Fig. 1). The hole was cored using a truck-mounted, cable tool, water well drilling system. The hole was cased using standard 8 inch water well steel casing, which was advanced behind the corer every 3 m. This eliminated significant down-hole contamination. The top 42 m of material was cored using a 1 m long, 75 mm internal diameter push corer mounted on the drilling bit. Below 42 m, samples were recovered using a 100 mm internal diameter hollow flight piston corer. These samples were recovered in 0.25–0.50 m increments and came out as cohesive core except in loose gravels near the bottom of the hole.

2.2. Sedimentary descriptions

All samples were described visually either in the field (bagged material) or the laboratory (core section). Attributes such as the presence of worm casts, organics, concretions and coarse lithics were recorded. Munsell colours were determined for all samples.

Depending on the visible grain size distribution, 20 g to about 250 g of material was extracted for particle size analyses, using a sedigraph for the finer than 63 μm fraction and an automated settling tube (Rapid Sediment Analyser) for the coarser material. The results were combined and composite grain size distributions and summary statistics using Folk and Ward (1957) were calculated. Grain size information is integrated into Fig. 2.

Based on the gross characteristics, informal sedimentary units were identified. Sixteen subsamples from these sedimentary units were recovered for XRD (X-ray diffraction) analyses of both the clay and, where appropriate, sand fractions. Samples for luminescence dating were selected on basis of sediment homogeneity and inferred origins (e.g. Berger, 1988).

2.3. Micropalaeontology

Samples were recovered for pollen, phytolith, diatom, and foraminiferal analyses. Initially sub-regu-

lar sampling at about 1 m intervals was undertaken. Foraminifera were found in only one sample and work in this area is unlikely to be pursued. Samples were examined at less than one metre intervals for both phytoliths and diatoms. Diatom preservation through the core was intermittent. Zones of good preservation occurred between -7 and $+1.5$ m AMSL and between -21.5 and -15.5 m AMSL. Some preservation also occurred between -63.5 and -57.5 m AMSL. From the initial examination, 50 samples were counted from these zones. A minimum of 300 diatoms per level were counted. Three intervals of phytolith preservation were recognised roughly matching those for diatoms.

Initial samples for pollen analyses were recovered at 1 m intervals but most proved barren. Subsequently, samples for pollen analysis were collected at irregular intervals from those sections of the core that offered the greatest concentration of organic material. Samples were obtained at relatively close intervals between -44.77 and -50.25 m. It is certain that important changes have been missed due to non-preservation of pollen. Damaged and unidentifiable pollen were found, and are believed to be reworked from the Late Cretaceous Gebbies Pass Plant Beds, presently exposed at the head of the valley (Sewell et al., 1992).

2.4. Luminescence dating

In ideal circumstances, sediments up to at least 800 ka old can now be aged by specific thermoluminescence (TL) procedures (Berger et al., 1994b; Berger, 1995). Therefore we chose twelve samples for TL dating, representing most of the recovered core length. These samples were extracted from the centre of the split cores (above -39 m AMSL) or the interior of large cohesive samples recovered during auguring (below -39 m AMSL). Parallel samples were recovered for dose-rate determination. Thick-source alpha-particle counting was used to measure U and Th concentrations, while K concentration was determined from atomic absorption spectrophotometry. Cosmic-ray dose rates were estimated and the average water content was assumed to be equivalent to the present saturation values for the samples. Alpha efficiency values (*b*-values) were determined from a comparison of extrapolated alpha-dose re-

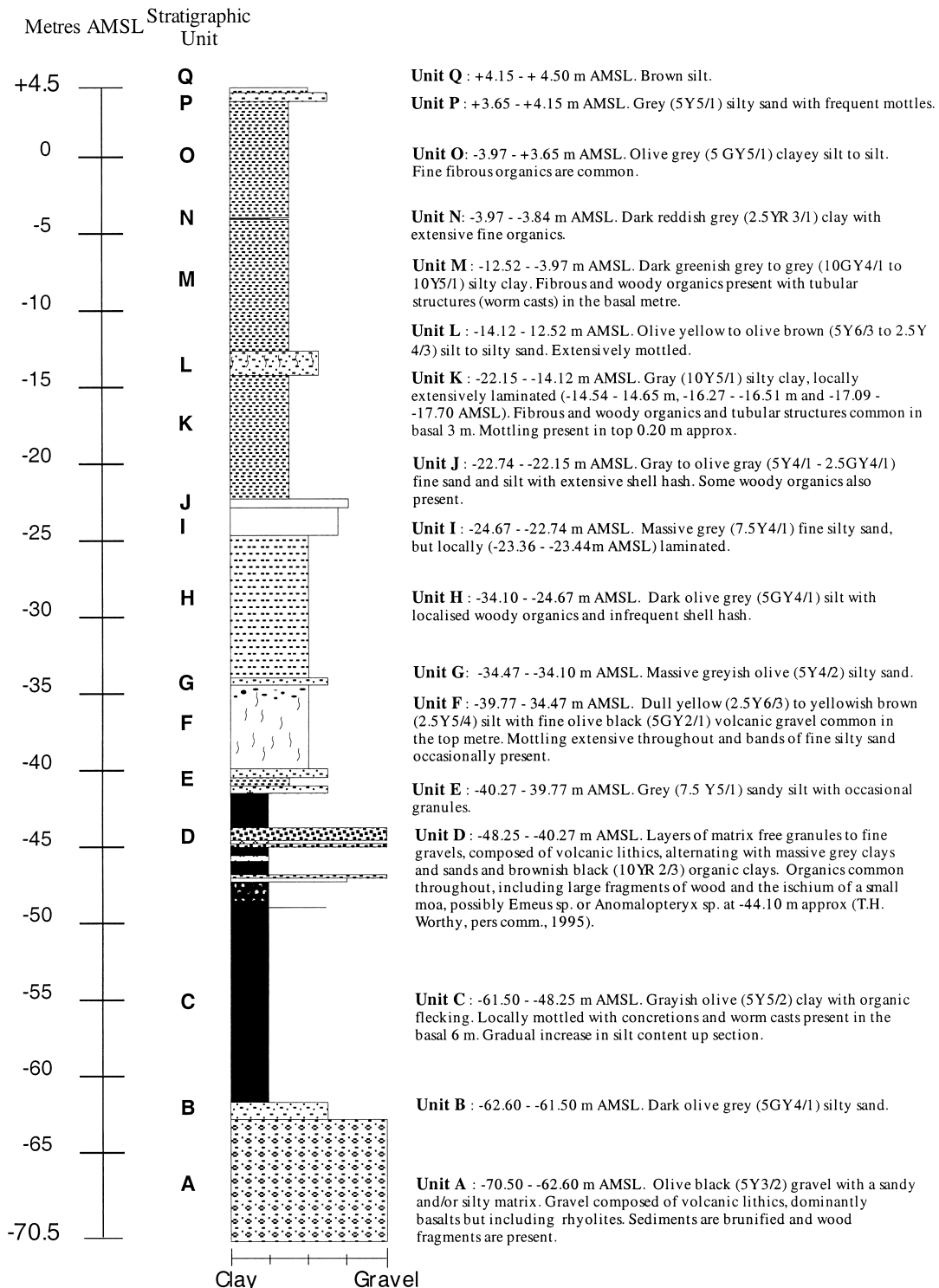


Fig. 2. Stratigraphy of the Gebbies Valley Borehole (G2).

Table 1
TL dating results

Sample ^a	Water ^b	K ₂ O ^c (%)	U (ppm)	Th (ppm)	<i>b</i> -value ^d (pGy cm ²)	<i>D</i> _R (Gy/ka)	Preheat ^e (°C)	<i>D</i> _E ^f (Gy)	Temperature ^g (Gy)	TL age (ka)
BP-27/11.48	0.42	3.32	2.18 ± 0.34	9.8 ± 1.1	0.84 ± 0.11	3.552 ± 0.096	120	32.6 ± 3.3 28.5 ± 2.7 38.1 ± 3.0	250–320/PB 250–340/TB1 220–280/TB2	9.18 ± 0.96
BP-25/17.38	0.24	2.28	2.59 ± 0.28	9.69 ± 0.92	1.07 ± 0.11	3.50 ± 0.13	130	85.8 ± 9.6 80.0 ± 5.4	230–310/PB 230–350/TB1	24.5 ± 2.9 2.8 ± 1.7
BP-21/19.06	0.44	3.10	3.06 ± 0.22	10.26 ± 0.73	1.04 ± 0.23	3.58 ± 0.17	140	413 ± 40 400 ± 64	220–310/ADD 220–340/PB	115 ± 12 111 ± 19
BP-23/21.90	0.38	3.28	3.21 ± 0.33	12.8 ± 1.1	0.73 ± 0.17	3.90 ± 0.16	140	365 ± 32 357 ± 69	240–320/TB2 250–340/PB	93.5 ± 9.1
BP-15/27.02	0.30	2.67	3.38 ± 0.37	8.4 ± 1.2	1.1 ± 0.3	3.71 ± 0.23	135	571 ± 68 570 ± 130	250–320/TB2 260–320/PB	154 ± 21
BP-13/32.68	0.34	3.01	2.82 ± 0.35	10.5 ± 1.2	1.1 ± 0.3	3.82 ± 0.22	140	434 ± 41 407 ± 59	230–300/TB2 230–340/PB	114 ± 13
BP-11/33.28	0.29	2.78	3.00 ± 0.34	9.7 ± 1.1	1.65 ± 0.26 (1.0 ± 0.2)	4.20 ± 0.22 (3.74 ± 0.18)	140	417 ± 40 417 ± 40 417 ± 58	230–300/ADD 230–300/ADD 240–350/PB	99 ± 11 (111 ± 12)
BP-9/42.47	0.28	2.18	3.48 ± 0.30	7.12 ± 0.98	1.47 ± 0.20	3.60 ± 0.17	135	429 ± 41 455 ± 66	220–380/TB2 220–320/PB	119 ± 13
BP-7/44.89	0.34	1.91	3.10 ± 0.31	6.0 ± 1.0	1.12 ± 0.16	2.79 ± 0.12	140	378 ± 25 348 ± 34	240–320/TB2 250–340/PB	136 ± 10
BP-5/55.62	0.37	2.28	3.30 ± 0.35	10.0 ± 1.1	0.71 ± 0.21	3.09 ± 0.17	130	555 ± 91 590 ± 160	200–290/TB2 210–300/PB	180 ± 31
BP-3/56.87	0.34	2.94	2.72 ± 0.37	8.8 ± 1.2	0.63 ± 0.14	3.28 ± 0.13	130	560 ± 110	210–300/PB	172 ± 33
BP-1/65.62	0.35	2.78	3.79 ± 0.34	11.4 ± 1.1	1.46 ± 0.48	4.31 ± 0.38	140	880 ± 140 830 ± 200	250–380/TB2 250–350/PB	204 ± 37

Please note that 4.5 m should be added to all core depth data to convert to m AMSL.

^a After slash: depth (m) within core.

^b Ratio mass water/mass dry sample. Uncertainty is ±0.02 for first sample and ±0.03 for others. Errors are ±1σ throughout table.

^c Uncertainty is ±0.05.

^d Alpha efficiency factor. This was not measured for samples 15 and 13, but only estimated from measured range for other samples.

^e For 4 days.

^f Weighted (by inverse variance) mean and average error over temperature range in next column.

^g Selected temperature range in *D*_E–*T* plot. The lower limit is 10°C below lowest data point because each point represents a 10°C slice of the readout curves. The *D*_E-measuring method is indicated after the slash. Partial bleach (PB), total-bleach1 (TB1), total-bleach2 (TB2) and additive dose (ADD) (Berger et al., 1994a). These methods are illustrated in the figures. TB2 used bleaching for 24 h with a full-spectrum Hg-vapour lamp. PB used 550–700 nm wavelengths for samples 3, 5, 21, 25, and 27 and TB1 used the 435–700 range for these samples. PB used the 435–700 nm range for the other samples, and TB1 used the range 550–700 nm for all of the others except sample 15. For age calculation, the *D*_E value having the lowest error was chosen, when comparing PB and either TB2 or ADD results, when these results agree with 1σ. For sample 27, the TB2 result is ignored (see Fig. 3C).

sponse curves with beta-dose response curves (see Berger, 1988). Alpha doses were administered by sealed-foil ^{241}Am sources. The above data were used to calculate a dose rate (D_r) (equations in Berger, 1988).

A luminescence age is equal to the equivalent dose (D_E) divided by the dose rate D_r . D_E values were determined from extrapolations of beta-dose response curves. Beta doses were administered by a 125 mCi sealed-foil ^{90}Sr – ^{90}Y source. The partial-bleach and total-bleach procedures (Aitken, 1985; Berger, 1988) were used to measure D_E values, which were then plotted against readout temperature to provide a plateau test. To remove thermally unstable and anomalous-fading TL components from the laboratory-irradiated subsamples prior to TL readout, elevated-temperature storage treatments were employed (see refs. in Berger, 1994).

The chosen preheating temperatures are listed in Table 1. Subsamples of polymineralic grains 4–11 μm in diameter were heated at 5°C/s. The TL was recorded through a 5-mm-thick KOPP optical glass filter CS-560, which transmits 360–495 nm wavelengths at 5% cut (345–505 nm at 1% cut). Thus most ultraviolet emissions are blocked, as are those from several plagioclase feldspars, and the detected signal is dominated by the ca. 410 nm emissions common to most K-rich feldspars (e.g. Berger, 1995).

2.5. Radiocarbon

Six samples were submitted for AMS dating at the Institute of Geological and Nuclear Sciences laboratory in Wellington, New Zealand.

3. Results

A summary stratigraphic column is presented in Fig. 2. Of the 75 m of material recovered, most (about 60 m) are fine silts and clays. Gravels are common only in the basal 10 m. In the clay fraction, XRD indicates that kaolinite, mica and chlorite are ubiquitous with smectite and vermiculite locally present.

3.1. TL dating

Age determinations from TL are presented in Table 1. The TL results indicate that about 12.5 m of Holocene muds and silts overlie a yellowish silt (loess) of Last Glacial Maximum age (BP-25: 22.8 ± 1.7 ka) (Fig. 3). The major feature of the TL results is the concentration of ages around 90–130 ka indicating a thick sequence of deposits of Marine Isotope Stage (MIS) 5 (and possibly 6) between –12.5 and –43.75 m. Of these ages, BP-15 at –22.52 m AMSL is anomalous because of its larger age and greater error range (though it still overlaps surrounding samples at 2σ). The basal three TL samples between –51 and –61 m yield ages averaging around 190 ka and may represent either MIS 7 or MIS 7 and MIS 8 deposits.

3.2. Radiocarbon dating

Radiocarbon dating results are presented in Table 2. Only one completely satisfactory age (NZA 5262) was obtained from a wood sample recovered at a depth of –7.08 m AMSL. This yielded an early Holocene age of 7558 ± 89 BP (8271 ± 89 CAL BP: Stuiver and Reimer (1993)) which is consistent with the results from the luminescence dating. Two of the remaining samples yielded finite ages. NZA 5264 (–21.54 m AMSL) yielded an age of $43,900 \pm 1700$ BP. This is at the limit of the radiocarbon technique (the sample has only 0.42% of the modern C-14 content) and is considerably younger than the TL ages from these levels. The age of $37,230 \pm 620$ BP derived from NZA 5663 at –54.75 m AMSL is clearly erroneous as there are three overlying radiocarbon samples which yield minimum ages that are significantly older and occur up to 30 m higher in the core. In addition, two consistent TL (BP-3 and BP-5) ages of around 180 ka were recorded from within a few metres of this sample.

3.3. Diatoms

Over 130 species of diatoms were identified from samples in the core. Of these about 30 species occurred commonly or in significant numbers in sections of the core. They were preserved in three zones. The lowest of the zones occurs from –63.5 to

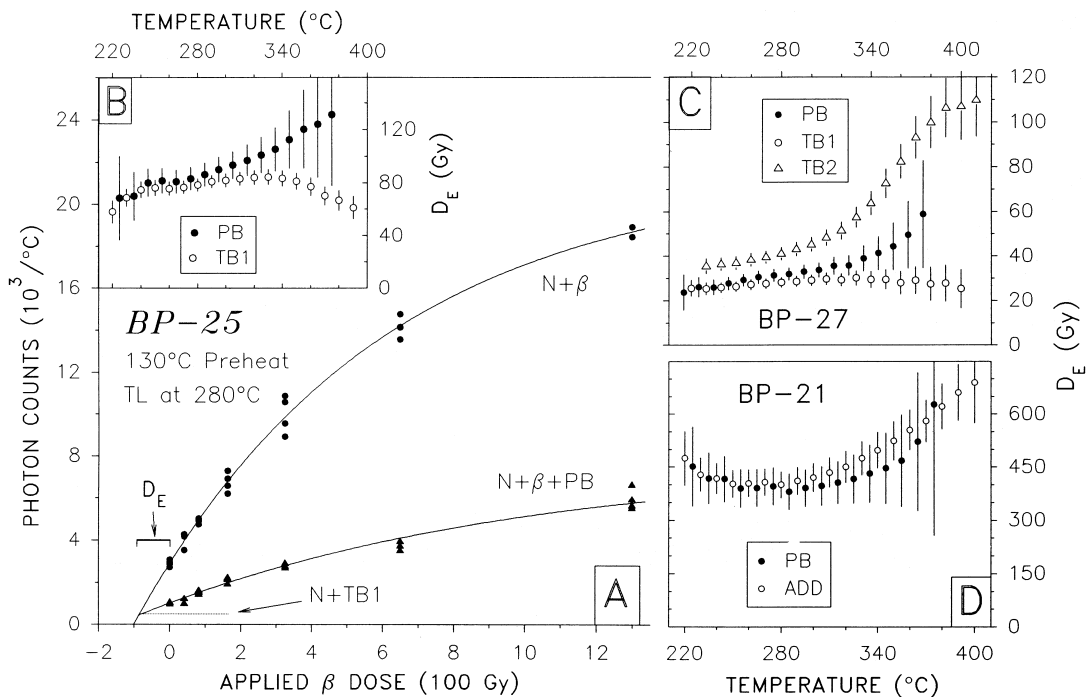


Fig. 3. Luminescence growth curves for samples BP-25 (A, B), BP-27 (C) and BP-21 (D).

–56.75 m AMSL. Only four samples were analysed from this zone as preservation was discontinuous. The diatom assemblages in these samples are presented in Table 3 and reflect a mixed freshwater and marine source. Some of the key taxa are presented in Plate I.

The next 30 m of sediment were barren. There is a single diatom sample at –26.8 m AMSL (not shown) and a major zone of diatom preservation which starts at –21.0 m AMSL approx (Fig. 4). Marine diatoms, of inferred coastal provenance, dominate the lower part of this section between –26.5 and –18.5 m AMSL. In the main, the marine diatom *Melosira*

westii is the dominant species accounting for 50–95% of all diatoms recorded at these levels. These samples have very limited floras and, in general, only small-size *Fragilaria* species and *Achnanthes delicatula* are consistently present with *M. westii*. The sole exception is a sample at –21.12 m AMSL with a mainly marine assemblage composed of *Diploneis parma* and *Thalassiosira eccentrica* along with some brackish and freshwater taxa including *Achnanthes delicatula*, and small *Fragilaria* species.

At –18.5 m AMSL there is an abrupt change to a freshwater assemblage dominated by small *Fragilaria* species (up to about 80%), with *F. brevistriata*,

Table 2
Radiocarbon (AMS) dating results from the Gebbies Valley core

Sample	Sample depth (metres below MS)	$\delta^{13}\text{C}$	Age (years BP)	Stratigraphic unit	Notes
NZA5262	7.08	–26.4	7558 ± 89	M	wood
NZA5264	21.54	–26.7	43,900 ± 1700	K	wood
NZA5376	22.66–22.70	–1.4	>50,000	J	shell hash
NZA5263	28.48	–26.4	>47,000	H	wood
NZA5244	42.70–43.10	–24.8	>47,000	D	wood
NZA5663	54.75–55.00	–26	37,230 ± 620	C	wood-contaminated sample

Table 3
Diatoms recorded from oxygen isotope stages 7 (and 8?)

Diatom species	Habitat	Depth (AMSL) and stratigraphic unit			
		Unit C (–56.75 to –57 m)	Unit C (–58.25 to –58.5 m)	Unit C (–58.75 to –59 m)	Unit A (–63.75 to –64 m)
<i>Cocconeis scutellum</i>	M			5.55	
<i>M. westii</i>	M				18.69
<i>Diploneis subovalis/smithii</i>	B	3.22	8.30		
<i>Gyrosigma wansbeckii</i>	B	7.91			
<i>Aulacoseira granulata</i>	F		5.53		
<i>Cocconeis placentula</i>	F	4.39	5.84		8.41
<i>Cymbella kappii</i>	F	7.33			
<i>Epithemia adnata</i>	F	28.44	19.69		47.66
<i>E. sorex</i>	F		12.30		7.47
<i>Fragilaria</i> sp.	F	7.91	7.69	11.1	
<i>F. ulna</i>	F	12.90	6.46		
<i>Melosira</i> sp.	F		13.84	77.77	
<i>Nitzschia fonticola</i>	F			5.55	

Only types recorded in at least one sample at above 5% are recorded. Environmental indicators: F = fresh water, B = brackish water, M = marine (from Van Dam et al., 1994; Harvey, 1996). Columns do not total 100% precisely because of the exclusion of taxa with trace values.

F. ulna, *Cocconeis placentula*, and a variety of *Cymbella* and *Epithemia* species. This change is reversed briefly at –18.06 m AMSL where *Melosira westii* is again the dominant species, in a very limited flora. Excluding this level, small *Fragilaria* species dominate to about –16.5 m AMSL, though they decline from 70–80% of the total at the base to 40–50% just below –16.5 m AMSL. At –16.00 m, a single sample is dominated by the freshwater taxa, *F. ulna*, *Cocconeis placentula*, *Cymbella affinis* and *Gomphonema minutum*. Between –16.00 and –15.50 m AMSL, *Aulacoseira ambigua*, *A. granulata* and *Cyclotella stelligera* become co-dominant. Above –15.50 m diatom preservation ceases.

Diatoms recur at –7 m AMSL (see Fig. 5). From –7 m to –3.90 m AMSL the assemblages are dominated by the marine and brackish types *Nitzschia granulata* and *Diploneis subovalis/smithii*. At –3.9 m there is an abrupt transition to a freshwater assemblage composed of small *Fragilaria* species. While the small *Fragilaria* species remain present throughout, there is a brief reversion to a marine assemblage at –1.8 m AMSL. This differs from the –7 m to –3.9 m saline assemblages in the dominance of *Melosira westii*. After freshwater conditions had been re-established at –1.7 m AMSL, *Aulacoseira* and *Cyclotella* species replaced the small *Fragilaria* species as the major elements in the assemblage.

Above +1.5 m AMSL diatoms disappear, probably as they are not preserved in the coarser sediment.

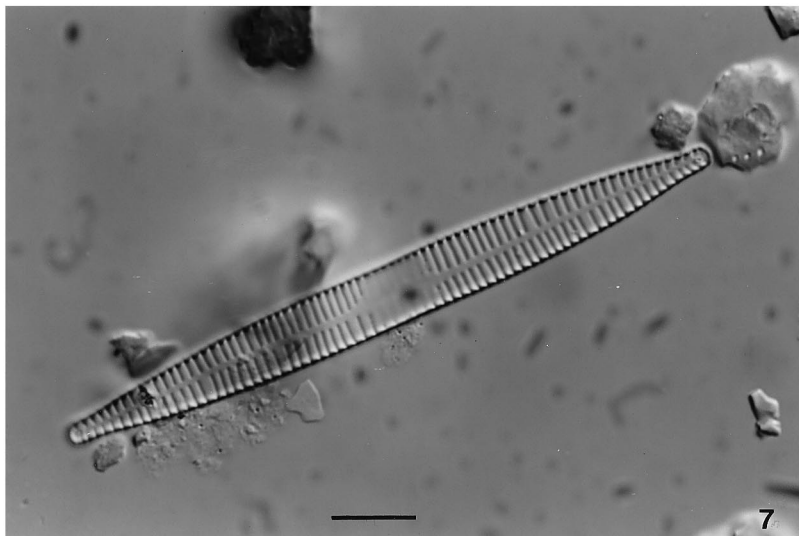
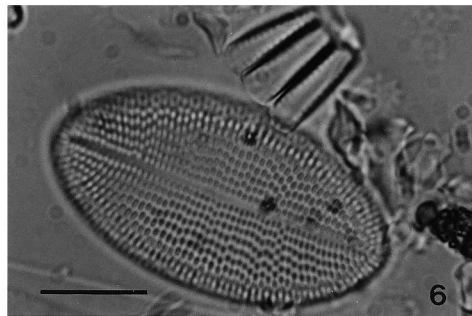
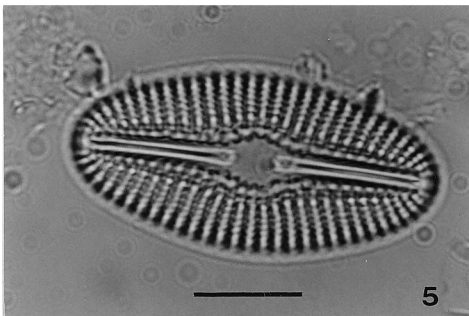
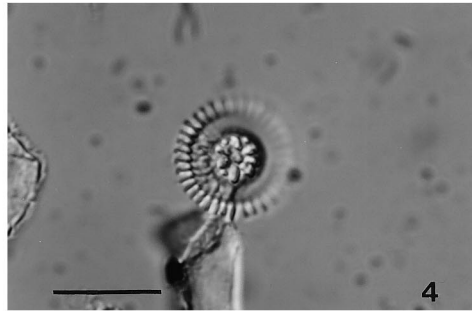
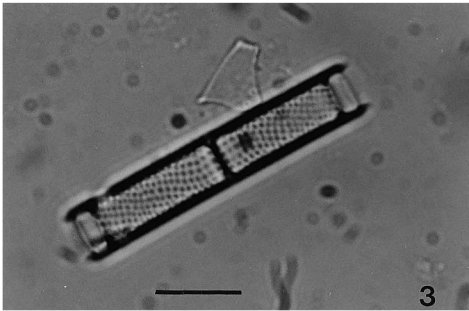
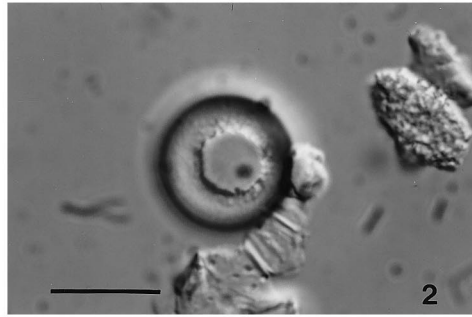
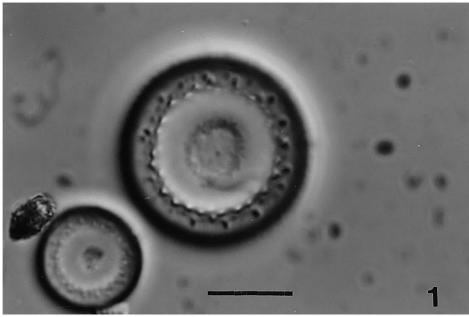
3.4. Phytoliths and pollen

The major phytolith morphological forms are presented in Plate II. Two zones of pollen and three zones of phytolith preservation are recognised. Summary results for pollen are displayed in Fig. 6 and for phytoliths in Fig. 7. Minimum counts of 300 microfossils were undertaken for both palynomorphs and phytoliths.

In the lowest zone between –60.5 and –56 m AMSL the phytolith assemblage is mixed with raised levels of tree (spherical) forms throughout and an increase in grass (chloridoid and chionochloid) forms at the top. Phytoliths recur between –14.5 and –21.5 m AMSL, where tree (spherical) forms dominate. Polyhedral forms are also common. An abrupt decline in spherical phytolith forms occurs at –16.5 m AMSL.

Between –14.5 and –5.5 m AMSL phytolith preservation was poor with only a few weathered polyhedral forms. Phytoliths are preserved between –5.5 to +4.5 m AMSL, where the assemblage is dominated by the grass forms (chionochloid and chloridoid) which peak briefly between +2.5 m and +4.5 m AMSL. In this zone spherical phytoliths

PLATE I



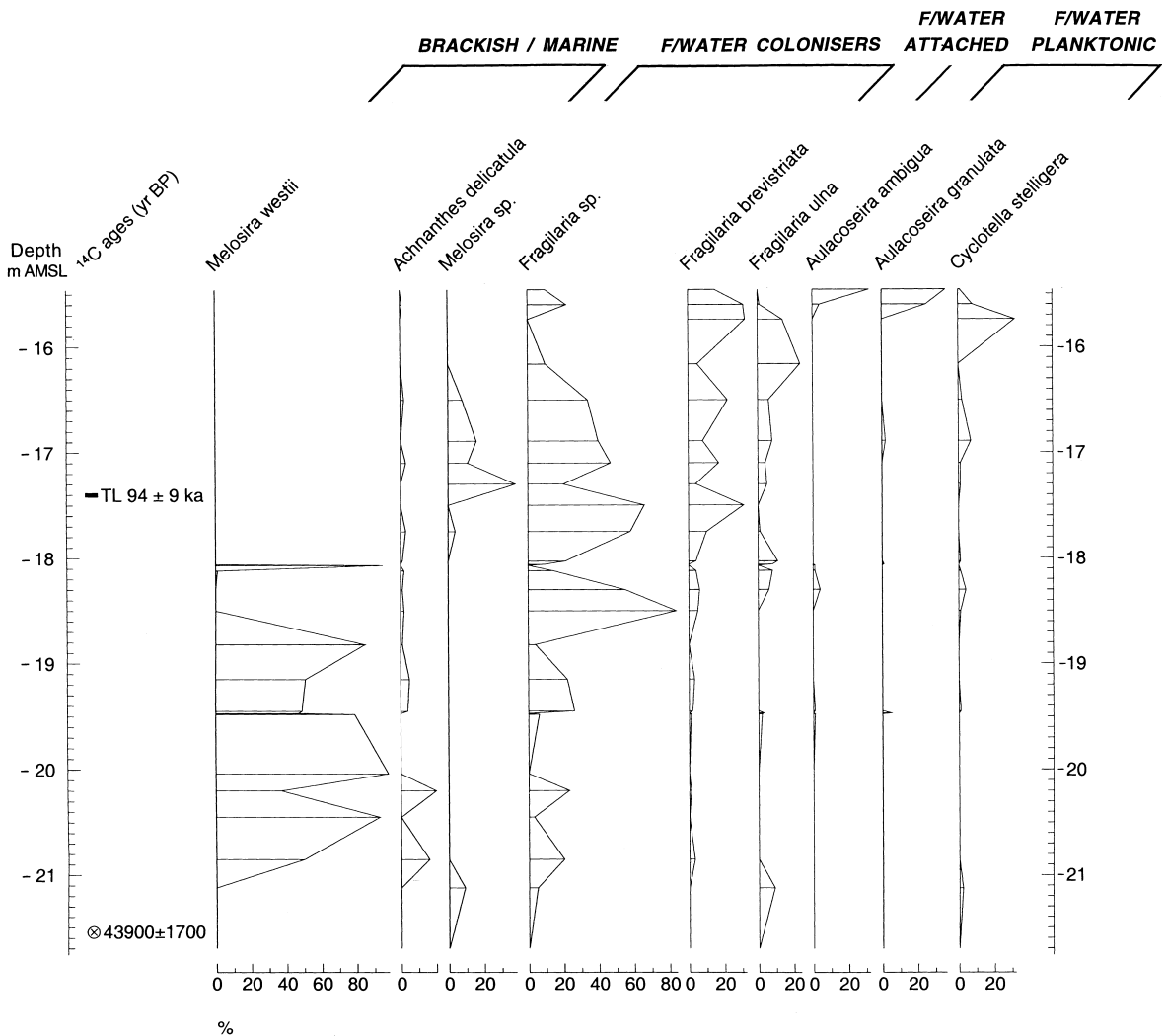


Fig. 4. Summary diatom diagram for Isotope Stage 5(a?). This diagram shows the percent abundance of the major indicator diatom species for marine, brackish and freshwater conditions and differentiates between attached and planktonic freshwater species. Note that species that occur in low values and/or species which occur at only a few levels have not been presented. Note also the succession from saline to freshwater attached/colonising taxa and then to freshwater planktonic taxa.

PLATE I

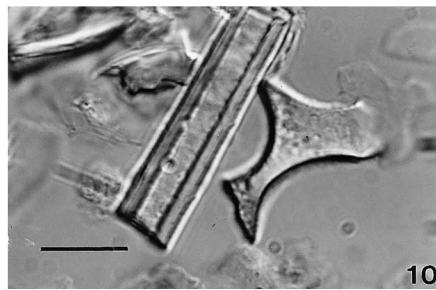
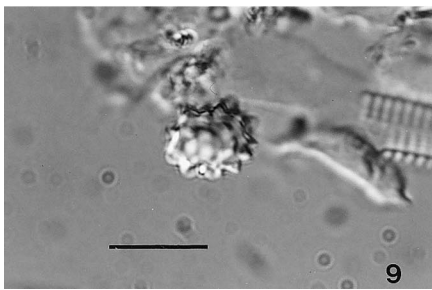
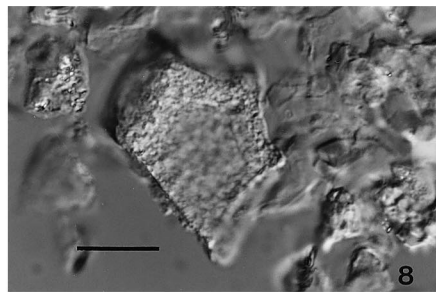
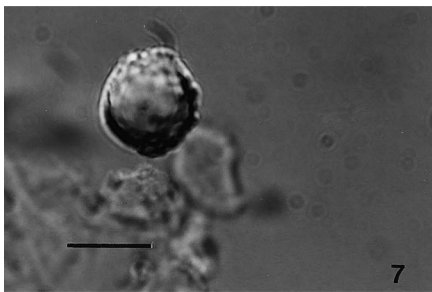
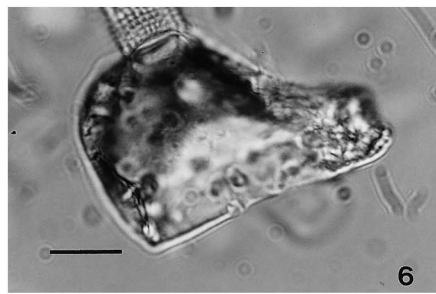
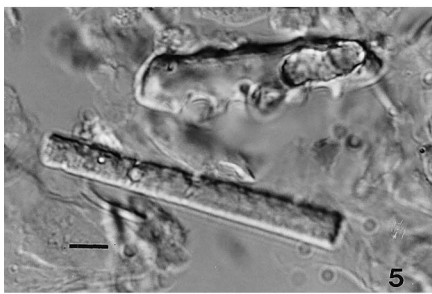
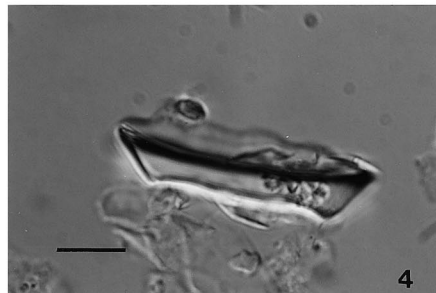
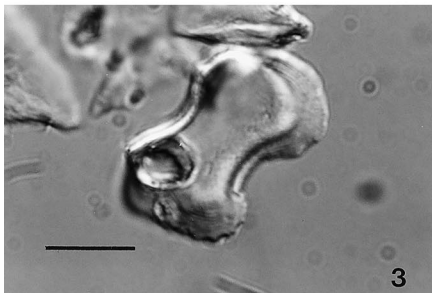
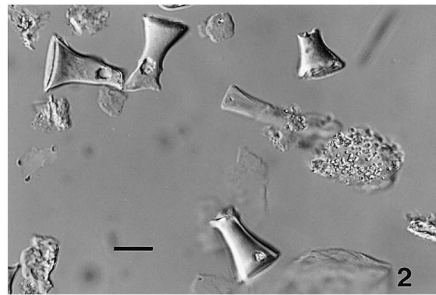
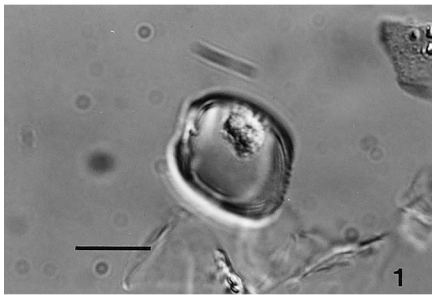
Plate showing some of the key marker diatom taxa in the core.

1. *Melosira westii*.
2. cf. *Melosira* sp.
3. *Aulacoseira ambigua*.
4. *Cyclotella stelligera*.
5. *Diploneis subovalis*/smithii.
6. Pseudoraphe valve of *Cocconeis placentula*; above it three linked frustules of small *Fragilaria* sp.
7. *Fragilaria ulna*. Scale bars are 10 µm.

are rare (<15%) but polyhedral forms are common (>40%) except near the surface, where both tree forms are rare.

Palynomorphs were preserved between -49 and -40 m AMSL and more discontinuously in the top 10 m. Podocarps dominate in the top 10 m, in a single sample at -58 m, and between -49 and -44 m AMSL. Between -44 m and -40 m, small trees and shrubs, notably *Leptospermum* type, *Coprosma* spp., and *Plagianthus*/*Hoheria*, are dominant.

PLATE II



4. Sediment sources in Gebbies Valley

The parent rocks in Gebbies Valley are Miocene basalts and a Cretaceous rhyolite with a very limited exposure of sedimentary rock. Consequently, one interesting feature of the sedimentary record in Gebbies Valley is the absence of clays that can be positively attributed to mafic rocks. There are, for example, no traces of iron oxyhydroxides (e.g. goethite) in the XRD results and all the clays recorded (kaolinite, smectite, vermiculite, chlorite and micas,) are compatible with a feldspar-rich greywacke parent material. This suggests that the clay fraction is derived substantially, though not necessarily uniquely, from greywacke. By contrast there is *no* greywacke gravel in the valley. In all samples with a gravel component, basalts and rhyolites dominate the gravel fraction.

Beyond the mouth of the valley, to within a few tens of metres, greywacke gravels are ubiquitous in the alluvial Canterbury Plains. Based on the absence of greywacke gravels, the distal ends of the fans never spread into Gebbies Valley. Furthermore, erosion of local rocks is a minor contributor to the fine sediment fraction, as shown by the absence of mafic indicators in the XRD analyses. A source of material with the appropriate size and mineralogical characteristics is present, however, in the form of the extensive loess deposits on valley sides. These valleys were also transgressed during high sea-stands.

We conclude that marine and aeolian processes were responsible for the transfer to, and accumulation of fines in, the valley bottom. This situation is comparable to that found in all the peninsula valleys running out to the Canterbury Plains, and to Lyttelton Harbour, where the silts which infill the harbour are considered to be derived from loess deposits (Brodie, 1955; Bushell and Tear, 1975; Curtis, 1985).

5. Environmental history

Unit A is interpreted as a terrestrial deposit. It displays evidence of pedogenesis based on colour and the presence of worm casts but the sediments are of fluvial origin. The one diatom sample was recovered about 0.9 m below the contact with the overlying unit. It contains a rather mixed assemblage composed mainly of diatoms with freshwater weed affinities (*Epithemia adnata*, *E. sorex*) but with about 20% marine and brackish types including *Melosira westii*. Our preferred explanation is that the marine types are present due to salt water incursion into the valley either because the river valley was tidal or at the start of a marine transgression. In either case the valley floor was at or very near sea level.

Based on sorting and colour criteria, Unit B is interpreted as an aqueous deposit. With the very limited information we have, it is not possible to distinguish between a marine or fluvial origin for these silty sands.

Quiet water deposition is indicated for the massive organic-rich clays of Unit C. Siliceous concretions are present, especially in the basal 3 m, and as a result diatom preservation is poor. The limited diatom flora preserved from –59 m to –56 m AMSL suggest a dominantly freshwater environment with a connection to the sea. Marine dinoflagellates were present in the basal member of the unit, confirming connection of the site with the open sea. The phytolith assemblage is dominated by spherical types normally attributed to trees (Kondo et al., 1994). The pollen indicates a mixed podocarp–broadleaf forest similar to the present (pre-European occupation) flora with *Prumnopitys taxifolia* (Matai) dominant and with some *Nothofagus fusca* type present at least regionally, in the basal sample. The pollen indicates full interglacial conditions for this unit. High values for

PLATE II

Major phytolith morphological forms.

1. Chloroid: Battle axe saddle-shapes (Tussock grasses).
2. Chionochloid: spool shapes (*Chionochloa* grasses).
3. Panicoid: dumbbell forms from 'warm' climate grasses.
4. Festucoid: elongate 'boat'-shapes with saw toothed edges (*Poa* and *Festuca* (grasses)).
5. Elongate: rectangular (grasses).
6. Bulliform: fan-shaped (*Rytidosperma* (grasses)).
7. Spherical verrucose: spherical with surface ornamentation (undifferentiated trees and shrubs).
8. Polyhedral epidermal: platey shapes with four to eight sides (mainly *Nothofagus* spp.).
9. Spherical spinulose: spherical with spines (palms; there is only one member of the family in New Zealand, *Rhopalostylis sapida* (Nikau)).
10. Jigsaw anticlinal: platey jigsaw shapes (undifferentiated trees, shrubs and ferns). Scale bar is 10 μ m.

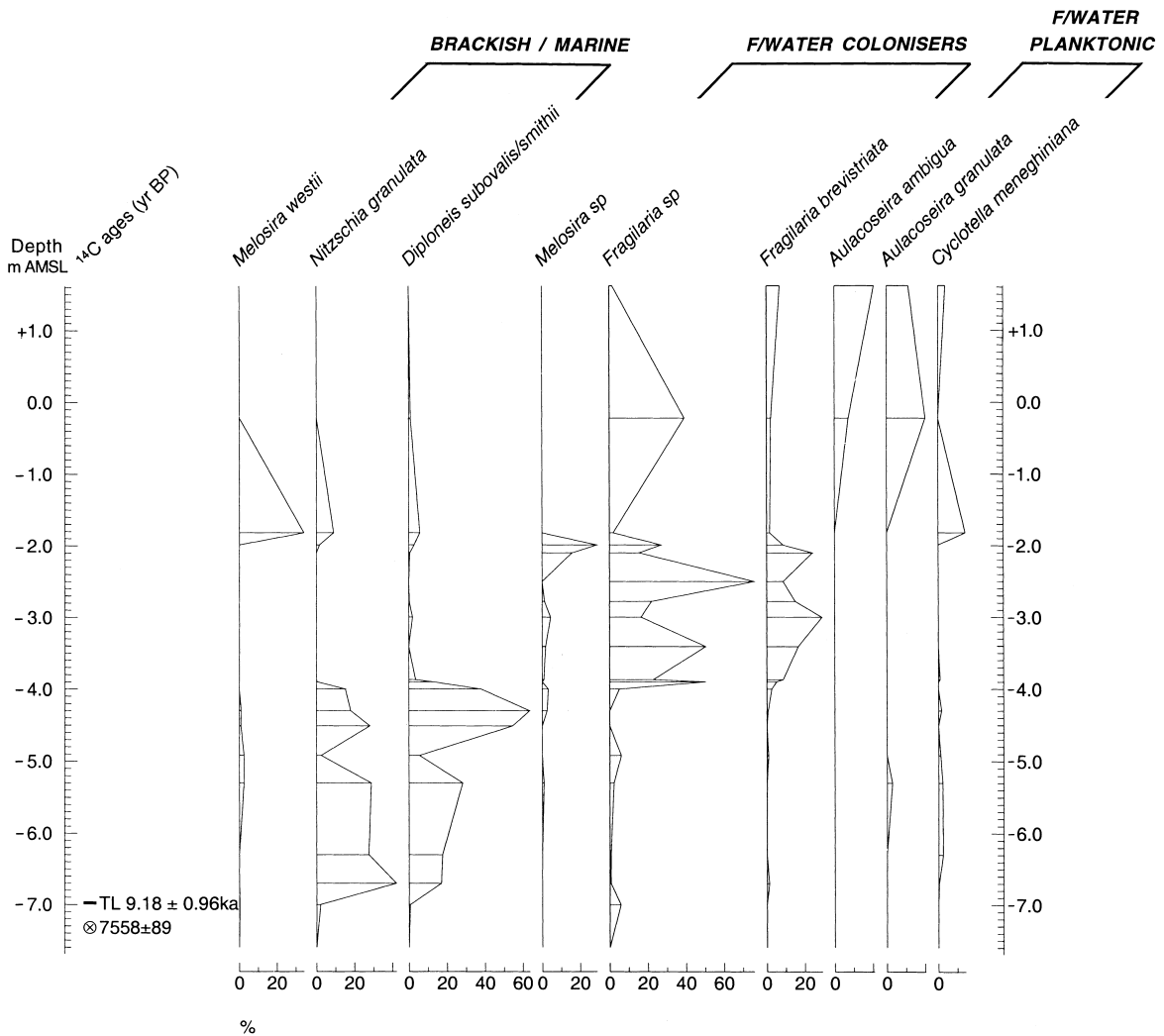


Fig. 5. Summary percentage diatom diagram for the Holocene. Note the overall similarity between the pattern in Holocene and that in Stage 5(a) (see Fig. 4).

raupo (*Typha*) pollen and the presence of *Glochidia* and spores of the water fern (*Azolla*), not shown on the pollen diagram, confirm the diatom interpretation of a quiet water environment such as a lagoon.

The major forest taxa decline at the base of Unit D and are replaced by a closed shrubland composed of *Phyllocladus*, *Plagianthus*, *Coprosma* and *Pseudopanax*. Manuka or kanuka (*Leptospermum scoparium*/*Kunzea ericoides*) dominated around the sampling site. This shrubland represents either a cool phase during an interglacial or, more likely, the onset of glacial conditions. The alternating beds of ma-

trix-free gravels, grey clays and occasional brown organic silts of Unit D are interpreted as barrier beach, shallow water and back swamp deposits, respectively, associated with a sea-stand at around -40 to -45 m AMSL. The presence of chenopods suggests local salt marsh environments which supports the sedimentological interpretations.

Above -40.27 m the thin sand (Unit E) may represent fluvial or marine conditions. This is overlain by 5 m of loess, soil and some terrestrial (fluvial) gravels (Unit F). Together the two units indicate a falling relative sea level to a low stand.

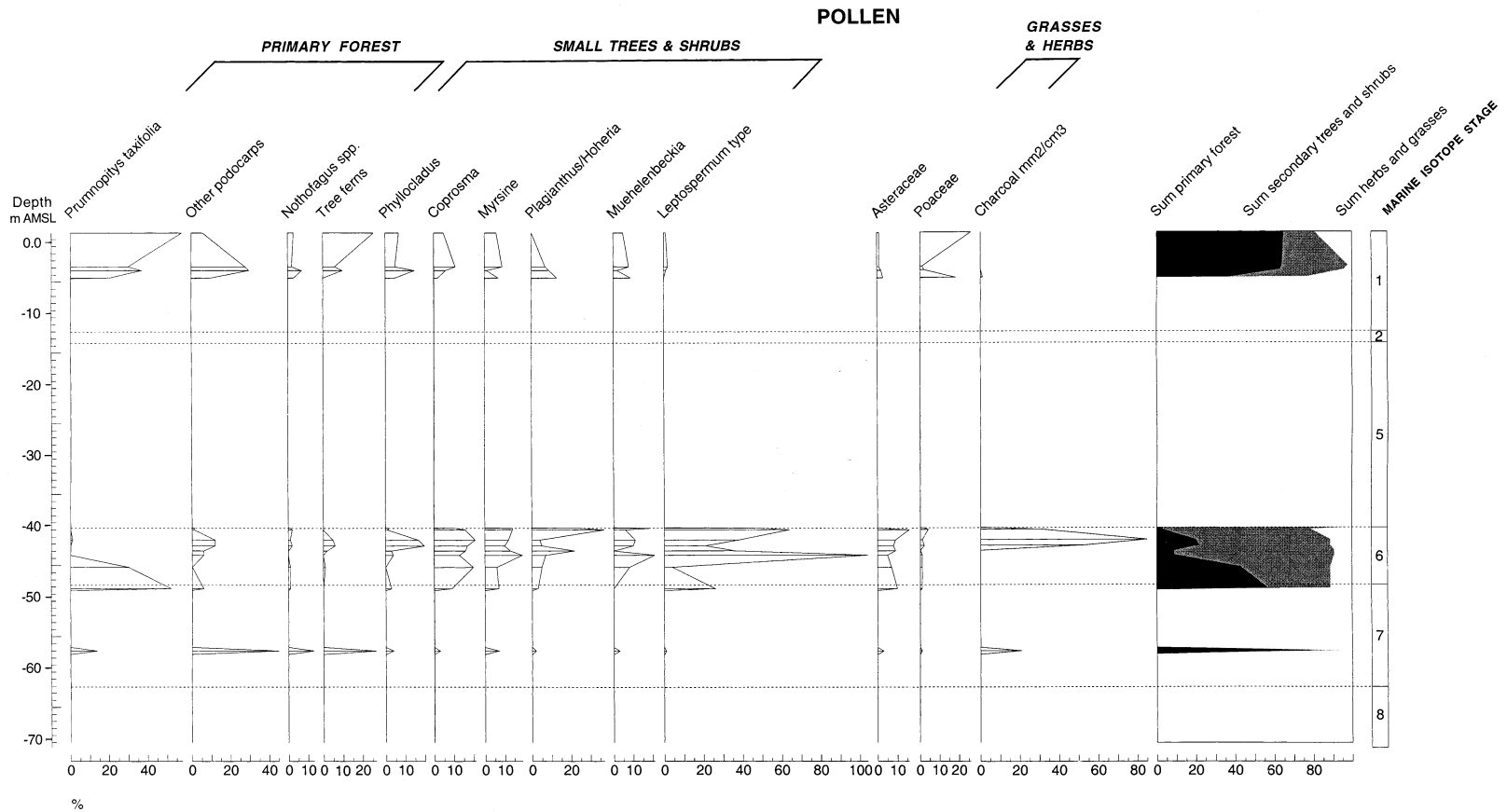


Fig. 6. Summary percentage abundance pollen information from the Gebbies Valley Borehole. The primary features are the presence of woody taxa throughout which indicates that at no time when pollen was preserved did the site lie above the tree line. The main cool climate indicator is the decline of primary forest taxa and its replacement with secondary forest and shrubland.

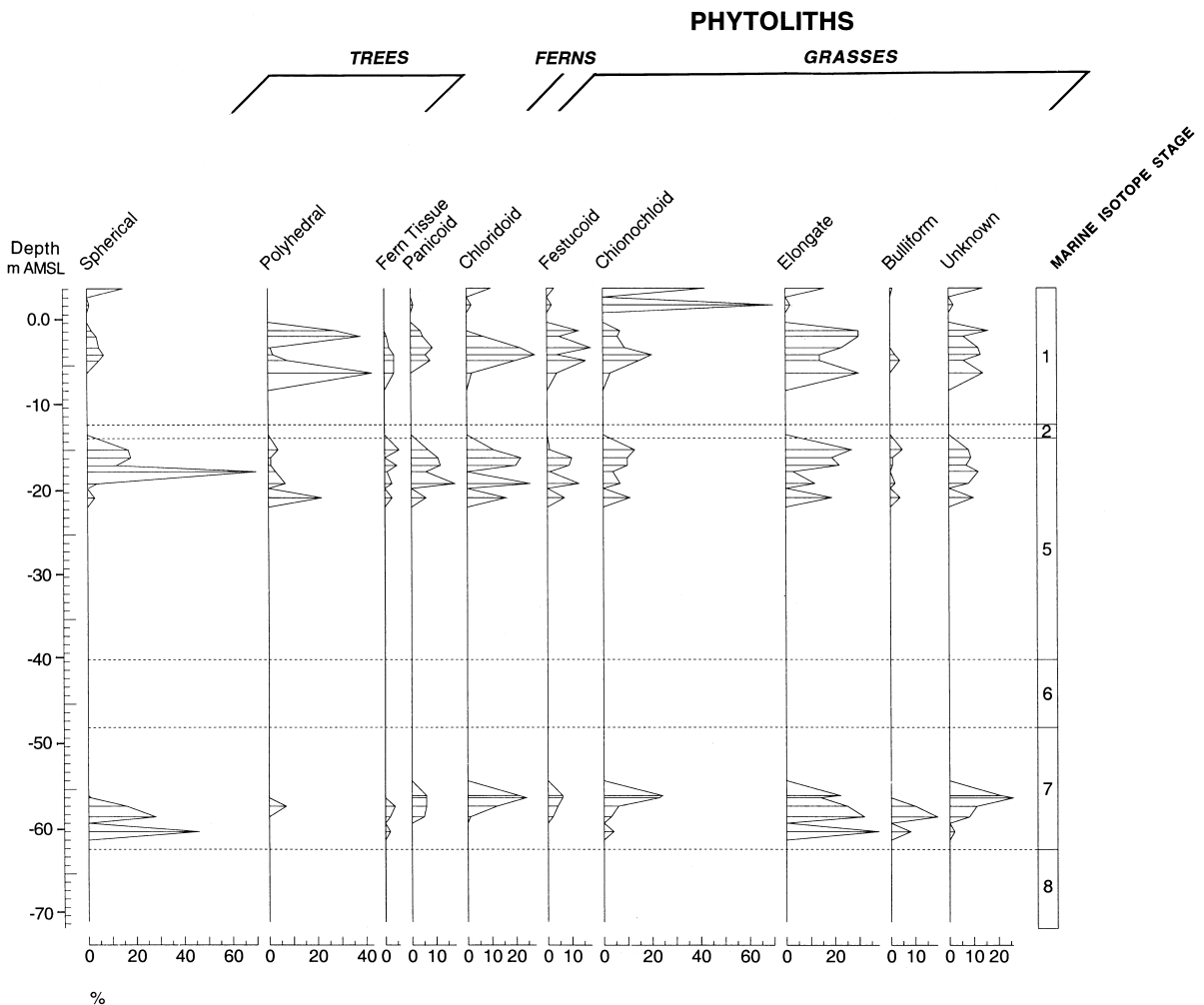


Fig. 7. Summary percentage abundance phytolith diagram. Note the very high values of the combined grass types throughout. Phytoliths are deposited more or less in situ, and the high grass values indicate the presence of grasslands around the edges of Lake–Lagoon Ellesmere, rather than a climatic effect. This is highlighted by the raised values of the chionochoid form in the top few metres, which is a response to grass occupation of the foredune ridge of the lake beach.

The terrestrial deposits are capped by a thin sand (Unit G) overlain by a thick sequence of grey silts (Unit H: -34.10 to -24.67 m AMSL). Shell hash is preserved, and where identified all shell material is of marine or estuarine origin. The marine origin is confirmed by a diatom sample at -26.8 m which is dominated by a single marine species, *Melosira westii* (90%). These deposits are therefore interpreted as a transgressive succession. The fine sands of Unit I (-24.67 to -22.74) are also interpreted to represent a relative sea-level fall.

Foraminiferal and macrofossil evidence from the fine sand and silt of Unit J (-22.74 and -22.15 m AMSL) indicates a beach or storm deposit. The *Ammonia beccari* fauna is intertidal and the shell hash which forms detrital layers contains cockles (*Austrovenus* sp.) which are of estuarine–nearshore provenance. Contemporary sea level was thus at, or very close to, -22 m AMSL. Beaches of Lake Ellesmere prior to the modern management of lake levels were not more than about 5 m above present sea level. This would give a minimum sea level

of about -27 m at Unit J time. However, if the Ellesmere embayment were not closed by a barrier, contemporary sea level would have been within 1–2 m of the level of the deposit.

On the basis of fine-grained sediment, Unit K represents a transgression and indicates a reversal of the shallowing trend in Units I and J.

Within the grey silts and clays of Unit K (-22 to -14 m AMSL) there are significant diatom assemblage changes from marine assemblages below -18.5 m, with a single recurrence at 18.06 m AMSL, to freshwater assemblages above -18.5 m. At -18.5 m there appears to be a transition from a marine embayment to a lake or lagoon environment. Initially, shallow (fresh) water colonisers such as the small *Fragilaria* species predominate but these are successively replaced by epiphytic species and then by planktonic freshwater species. The presence of Nikau palm (*Rhopalostylis sapida*) phytoliths in this unit are an indicator of probable interglacial conditions, as Banks Peninsula is the southernmost limit of Nikau distribution during the current interglaciation. Mottling and other evidence of pedogenesis in the top of the grey clays at -14.12 m AMSL indicate that the lake–lagoon sediments became exposed after sea level fell again.

Unit L is a mottled yellow clay and is clearly terrestrial. Grain size data, colour and pedological characteristics, all support an origin as loess. The upper contact of this unit is erosional.

Based on the diatom flora and the contact between aqueous and terrestrial sediments, Unit M (-12.52 to -3.97 m AMSL) clearly represents a marine transgression. There is no diatom preservation in the basal few metres but from -7.00 m upward, the flora is dominated by the marine diatom *Nitzschia granulata*, and the brackish *Diploneis smithii* with subsidiary counts of *Diploneis parma* (marine), *Nitzschia navicularis* (brackish) and *Fragilaria ulna* (fresh).

We interpret this as an estuarine environment. Pollen counts from this and the succeeding Units N and O are dominated by *Prumnopitys taxifolia* (matai) and other tree podocarps, indicating full interglacial conditions.

The red silty clay of Unit N is of very limited thickness (0.13 m) but marks an important change in aqueous sedimentation. A single diatom sample records an abrupt change to a freshwater assemblage

dominated by small *Fragilaria* species, *Navicula vulpina* and *Fragilaria capucina*.

The grey clays of Unit O represent lake–lagoonal deposits. There appear to be systematic changes in the diatom assemblages from a small *Fragilaria*- and *F. brevistriata*-dominated assemblages up to -1.8 m AMSL, through a brief reversion to marine conditions where *Melosira westii* is common (at -1.8 m AMSL), to an *Aulacoseira*-dominated (*A. ambigua*, *A. granulata* with *Cyclotella stelligera* and *C. meneghiniana*) assemblage above this point to about $+1.5$ m AMSL where diatom preservation ceases. This progressive change parallels the changes observed earlier in Unit K and is discussed below. The pollen record indicates the dominance of matai (*Prumnopitys taxifolia*) with high counts of tree ferns (*Cyathea* spp.). The pollen assemblage is consistent with the vegetation preserved in reserves on the peninsula today and represents an interglacial flora.

Unit P is the stratigraphic expression of the fore-dune ridges on the beach ridge of the historical Lake Ellesmere (Soons et al., 1997) and this is consistent with the high values of grass phytoliths found in this unit. Unit Q represents aeolian dust accumulated on the beach ridge since the abandonment of the lake beach due to artificial control of the lake level.

6. Chronostratigraphy and correlations with the Marine Isotope Stages

Age control for the core is derived from radiocarbon and TL dating. There is limited overlap between the radiocarbon and TL chronologies. The TL age from -7.04 m AMSL and the radiocarbon (AMS) age from -7.08 m from the Holocene section of the core comfortably overlap at two standard deviations. Also, the TL determination of a Last Glacial Maximum age (22.8 ± 1.7 ka, BP-25) from the underlying loess is consistent with the stratigraphy. Below the uppermost loess, radiocarbon dating fails and the chronology is provided from TL alone. Berger and Easterbrook (1993) and Berger and Anderson (1994) have demonstrated procedures that can produce reasonable TL chronologies from lacustrine deposits in excess of 100 ka.

The TL results are in general satisfactory, but to refine our age estimations, we have used palaeoenvironmental

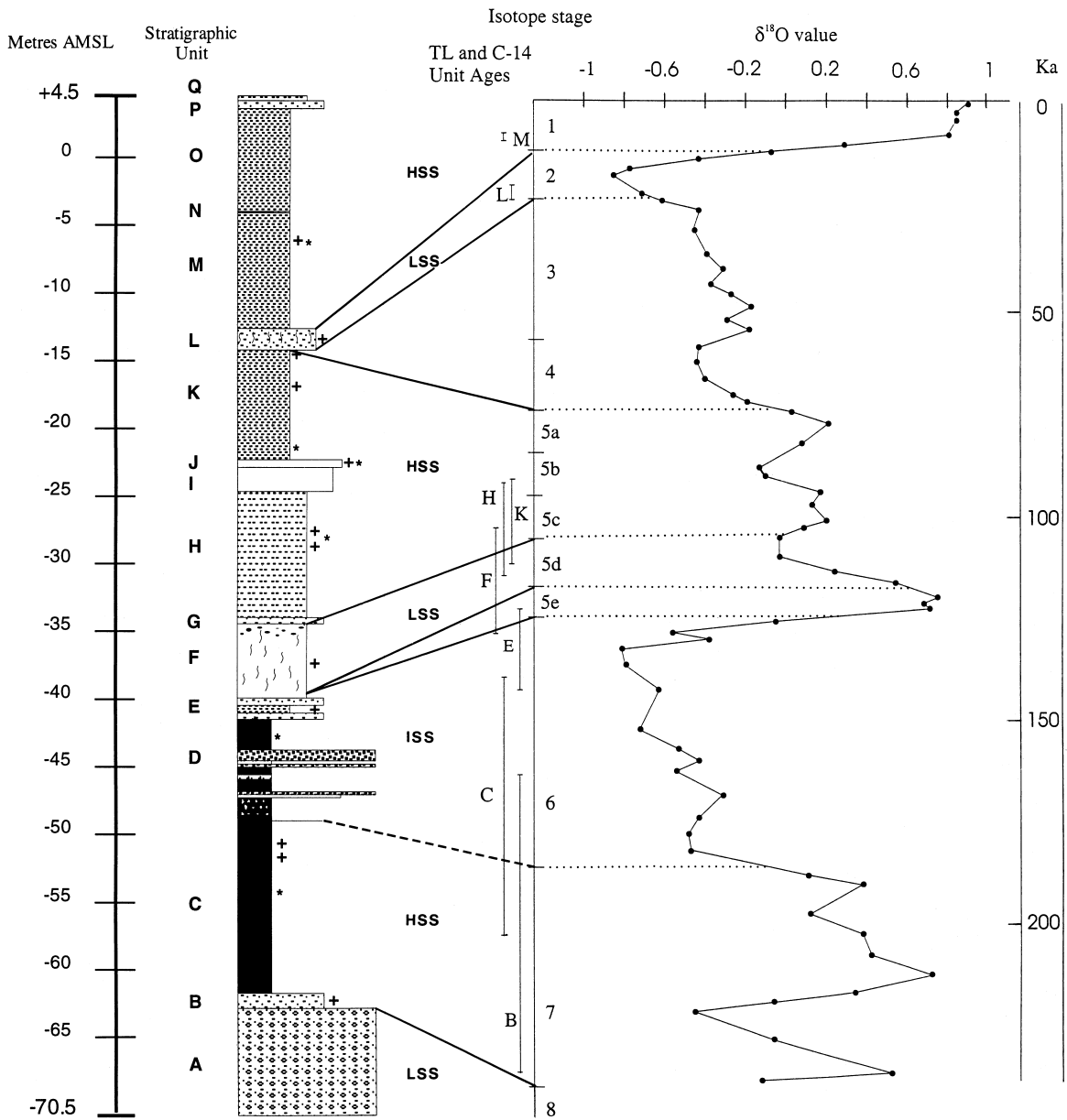


Fig. 8. Chronostratigraphy of Gebbies Valley Borehole and correlation to the orbitally tuned oxygen isotope chronology of Martinson et al. (1987). Sites of individual age samples are marked with an asterisk for radiocarbon samples and with a plus sign for TL samples (see Tables 1 and 2 for individual results). Age ranges for the units have been obtained from root mean square averaging of ages within single stratigraphic units. Unreliable ages (see text) have been excluded. Litho- and biostratigraphic information has been used to refine the chronology. Hence, Unit K which is virtually indistinguishable from Unit H on age estimates, is assigned to Isotope Stage 5a instead of also to 5c because the litho- and biostratigraphy imply a regression and subsequent transgression between Units H and K. Abbreviations: *HSS* = high sea-stand; *LSS* = low sea-stand; *ISS* = intermediate sea-stand.

ronmental data along with the TL ages, to calibrate our sedimentary units to the Marine Isotope Stages. Sea-level-corrected isotope curves are available for the last 125 ka (e.g. [Chappell et al., 1996](#)) but the precision of ages and sea levels for $\delta^{18}\text{O}$ stages older than Stage 5 is poorer than that for the younger stages ([Chappell and Shackleton, 1986](#)). Here we have chosen to calibrate our data against the standardised oxygen isotope curve for the last 300 ka ([Martinson et al., 1987](#)).

At the base of the core, Unit A records a dry valley floor, and sea level below about -60 m. This implies a cool to cold phase, and together with the TL age of 204 ± 37 ka suggests correlation with MIS 8 or a cool period in Stage 7 (e.g. MIS 7.4) (see Fig. 8). It is followed by transgression and eventual sea-level stabilisation in Unit C, with a forested environment consistent with full interglacial conditions, except in the top few metres where a cooling is indicated. The two TL ages recovered from Unit C are consistent. Both indicate ages of about 180 ka. Hence this unit is assigned to MIS 7 (the penultimate interglacial).

The sea level of -40 to -45 m AMSL inferred from the beach and related deposits of Unit D implies a relative sea-level fall from that in which the massive organic-rich clays of Unit C were deposited. The organic silts of this unit are likely to be freshwater deposits trapped behind a barrier beach, and as such were unlikely to have been more than 0–2 m above contemporary sea level. A cool phase is indicated from the pollen, and given the TL age of 136 ± 10 ka at the top of this unit, a temporary still-stand during MIS 6 is probable.

Above -40.27 m Units E and F indicate a falling relative sea level to a low stand. The loess of Unit F yields a TL age of 119 ± 13 ka, suggesting an association with a cool stage in MIS 5 or the peak of MIS 6. The obvious correlation here is with MIS 6. However, studies of North Island loesses report a period of deposition in MIS 5d (e.g. [Pillans, 1988](#); [Berger et al., 1994b](#)). Loess deposited at ~ 110 ka has also been documented in the South Island, near the mouth of the Awatere Valley in Marlborough ([Berger and Pillans, 1994](#)) and at the Clarence River ([Ota et al., 1996](#)). [Froggatt \(1988\)](#) considers a loess at Clifford Bay in the north of the South Island may also be of 5d age. Most of these papers also

document Stage 6 loesses. Uncertainty over the age of our Unit F does not affect the interpretation of the ages of the overlying units.

The terrestrial deposits of Unit F are capped by a thin sand (Unit G) overlain by a thick sequence of grey silts (Unit H: -34.10 to -24.67 m AMSL) which represent a marine transgression, with a sea level higher than -24 m AMSL. A TL result from -28.2 m AMSL shows an age of 114 ± 13 ka and a second sample from -28.7 m an age of 99 ± 11 ka. Both support a MIS 5 age. The latter age was acquired using an anomalously high b -value for these samples. With a 'normal' b -value (~ 1) for this sample (see Table 1), the age would recalculate as 111 ± 12 ka. Thus we prefer the older age of 114 ± 13 ka for this unit. This age is compatible with a MIS 5e or 5c interpretation. We note that full interglacial conditions associated with marine sedimentation in MIS 7 resulted in clay deposition, while the Holocene deposits are back barrier silty clays. The coarser texture of Unit H and the absence of interglacial pollen and dinoflagellates which should be preserved in deeper water deposits all tend to suggest a MIS 5c rather than a MIS 5e age for these sediments.

The fine sands of Unit I (-24.67 to -22.74) are tentatively interpreted as marine in origin. They represent a period of shallowing (possibly cooling) during MIS 5 (b?).

Unit J (-22.74 to -22.15 m AMSL) represents a beach or storm deposit, and as such would have been very close to sea level of the time. The juxtaposition of this sand between a coarse silty sand below and a fine silty clay above allows for this unit to either represent a continued shallowing from Unit I or more probably, a new transgressive deposit under the fine aqueous deposits of Unit K. This unit was beyond the limit of radiocarbon dating and the TL age of 154 ± 21 ka is both older and less well resolved than the ages above and below it. The sediments of Unit J may be the products of a storm event, or even a tsunami, and if so the light-sensitive TL could have been incompletely zeroed. This age is not statistically different from those around it, however. We suggest correlation with the rising limb of MIS 5a for the age of this unit.

The diatom assemblages of Unit K (-22 to -14 m AMSL) appear to record a transition from a marine

embayment to a lake or lagoon environment. A similar sequence recurs in the Holocene and it appears that a repetitive pattern of lake–lagoon formation has occurred from time to time along this section of the Canterbury coast. The TL ages of 94 ± 9 and 115 ± 12 ka indicate another Stage 5 unit. The younger age is compatible with a MIS 5a age and on stratigraphic grounds it must be either MIS 5a or less probably 5c. The upper contact appears unconformable as mottling and pedogenesis are present in the top of the grey clays at -14.12 m AMSL. An unknown thickness of material may have been removed.

Unit L is clearly a loess, and the TL age of 22.8 ± 1.7 ka corresponds with ages for the peak of the Otiran (Last Glaciation — MIS 2–4) derived from other studies (Suggate, 1993). This part of the core has been examined for evidence of the Kawakawa Tephra–Aokautere Ash (an important New Zealand stratigraphic marker at 22,600 yr BP), which could be expected at this level. It is known elsewhere in Canterbury (Campbell, 1986) but has not been certainly identified anywhere on Banks Peninsula. It may have been removed by the subsequent marine transgression in Unit M.

The TL age of 9.18 ± 0.96 ka, the C-14 age of 7558 ± 89 yr BP at -7 m near the top of Unit M, and the ages from Unit L all permit the upper 19 m of the core to be confidently assigned to the late Last Glaciation and the Holocene (Fig. 8). Unit M records the rise in sea level of the Late Glacial–Holocene transgression, and the overlying sequence to events relating to the building of the Kaitorete barrier and the isolation of Lake Ellesmere and Gebbies Valley from marine influences. These events have been described in some detail in Soons et al. (1997), and are now shown to have been accompanied by the establishment of a vegetation cover which was probably still present at the time of the first human settlement of the area, and even survived until the arrival of Europeans.

In summary, thick sequences of aqueous sediments associated with interglaciations (Isotope Stages 1, 5 and 7) are preserved in Gebbies Valley while glacial periods are confined to thin loesses and rare non-terrestrial sediments (e.g. Unit D). The absence of extensive cooler climate deposits may indicate non-deposition in sub-aerial environments during these periods, but at least four transgressive events are recorded in

the core and these may be responsible for truncating pre-deposited glacial age sediments.

7. Interglacial and interstadial vegetation of Banks Peninsula, lowland Canterbury

Despite the lack of continuity, the core offers significant insights into the interglacial and interstadial flora of lowland Canterbury. Summary pollen and phytolith diagrams are presented in Figs. 6 and 7. Of the taxa recorded in the core, only one, *Phyllocladus*, is not found on the peninsula today (H. Wilson, pers. commun., 1996) and this may be the result of a recent local extinction as it is present in Holocene pollen records. There is no evidence of glaciation on the peninsula (Soons, 1979), and while there may have been some periglacial activity on the highest areas (Harris, 1983) the deep valleys and varied aspects would have provided a range of protected microclimates where a wide range of species could survive even adverse conditions. The absence of *Phyllocladus* today is unsurprising as it predominantly a pioneer taxon and was probably outcompeted during the Holocene by the major forest taxa.

The pollen record from the Penultimate Interglaciation (MIS 7: Unit C) is dominated by *Prumnopitys taxifolia* (Matai), and other undifferentiated tree podocarps in association with understorey trees (*Coprosma* spp., *Myrsine* sp.). Interestingly, the highest values of southern beech pollen (*Nothofagus fusca* type) occur in association with high values of podocarps in the lowest pollen recording stratum. The presence of *Nothofagus* may indicate slightly cooler or wetter conditions in the lowest parts of the unit. Moar and Mildenhall (in Brown and Wilson, 1988) record a similar vegetation assemblage dominated by *Prumnopitys* sp. with *Coprosma* spp., and *Leptospermum/Kunzea* (manuka/kanuka) from the Shirley Formation (Penultimate Interglacial) in western Christchurch city. Phytoliths were preserved in the older sediments in MIS 7. The phytoliths record the local dominance of grasses, as might be expected on the fringes of an embayment or lagoon, with a strong background of tree types.

Our most detailed pollen records come from the Penultimate Glaciation (MIS 6: Unit D). These flora do not represent pleniglacial conditions which would

be recorded in the loesses if they were palyniferous, but interstadial conditions. This assemblage is dominated by secondary trees and shrubs, notably *Coprosma* spp., *Myrsine* sp., *Plagianthus/Hohera* spp. *Muehlenbeckia* sp. and *Phyllocladus* sp. *Phyllocladus* is the most notable as it is absent from the peninsula today and has a generally more montane distribution. The combination of coastal affinity plants such as *Plagianthus* with montane plants like *Phyllocladus* is unusual but not novel. That this site does not contain the pleniglacial flora is confirmed by comparison with the inferred Stage 6 flora from coastal Taranaki in the western North Island (Bussell, 1990). There, Bussell records a markedly more alpine flora dominated by grasses with daisies and montane shrubs.

In association with the coastal–montane shrubland, high values of charcoal and of the pioneer taxa, *Leptospermum/Kunzea*, indicate a rather unstable environment with frequent burning. In the context of the local climate, in which a dominant foehn wind over the Southern Alps has a strong drying effect, and controls the aeolian geomorphology of the Canterbury coast (Shulmeister and Kirk, 1996), it is tempting to attribute the burning to enhanced wind speed, and hence drying.

We have no pollen record from MIS 5 but phytolith data (Fig. 7), particularly the presence of Nikau Palm, suggest that local conditions were similar to both MIS 7 and MIS 1. Values for tree phytoliths indicate forest cover nearby while the records of grass phytoliths are consistent with swampy meadows, reeds and rushes along the embayment–lake edge.

The Holocene section of the core yields a palynoflora dominated by *Prumnopitys taxifolia* and other forest podocarps. Understorey shrubs of species such as *Coprosma*, *Myrsine*, *Plagianthus* and *Phyllocladus* remain important. There are minor floristic differences between this and MIS 7 but the floras are very similar. The phytolith data confirm the pollen patterns but show a strong rise in grass in levels above the uppermost pollen samples. This change is consistent with the development of a lake beach at the core drilling site during the late Holocene (Soons et al., 1997).

This intermittent, lowland terrestrial record can be directly compared to a long marine record (DSDP Site 594) recovered from the Canterbury shelf

(Heusser and van de Geer, 1994). The DSDP record displays alternations between periods of raised values of closed forest taxa (*Podocarpus*, *Nothofagus*, *Coprosma* and tree ferns) during interglaciations with periods of raised grasses and Cyperaceae during the glaciations. Thus our markers for interglacial flora match those of Heusser and van de Geer (1994) reasonably well, whereas our glacial indicators diverge with many more shrubs and small trees recorded at our site than in the marine record. This reflects both the absence of pleniglacial flora from our site, and its low altitude, where arboreal taxa probably persisted during stadials.

Our records differ from those of Heusser and van de Geer quite markedly during MIS 7 in that they have high *Libocedrus* and *Halocarpus* values from which they suggest a cooler interglacial by up to 3°C whereas our record suggests little difference between Stages 1 and 7 with high lowland podocarp values in both. This may reflect real discrepancies in the records but may be explicable simply in terms of pollen source, with the DSDP site receiving its montane flora from upland catchments via long-distance fluvial transfer and Gebbies Valley recording the local lowland flora.

8. A long-term sea-level history

The Gebbies Valley core provides an opportunity to develop sea-level information for this region of the southwest Pacific. Not only does the core contain evidence of repeated transgressions and regressions but the record is also unusual for a micropalaeontologically rich core, in that it is not very organic. Most of the sediments recorded are inorganic silts and much less compressible than peats or gyttjas. Even the most organic material at around –41 m AMSL is an organic-rich silt rather than a truly organic deposit. Compaction effects are therefore likely to be limited. In addition, unlike most areas of New Zealand, an assumption of tectonic stability during the late Quaternary, particularly the last few hundred thousand years, is consistent with the available evidence (Lawrie, 1993; Bal, 1997).

A summary of sea-level changes deduced from the core record is displayed in Table 4. While values for some points in the sequence are inferred, e.g.

Table 4
Summary sea-level data from core G2

Depth m AMSL + error estimates	Age (yr BP for ^{14}C , ka for TL and other)	Source of sea-level information	Sea-level data reliability
–6.98 + 0/–4	9.18 ± 0.96 ka	estimated max. depth of marine embayment	moderate
–7.08 + 0/–4	7558 ± 89 yr BP	estimated max. depth of marine embayment	moderate
lower than –12.88	22.8 ± 1.7 ka	loessic soil	directional information only
–14.56 + 4/–0	115 ± 12 ka	barrier blocked lake	high
–17.4 + 4/–0	93.5 ± 9.1 ka	barrier blocked lake	high
–20.18 + 0/–2	>50,000 yr BP	littoral sand	very high
–28.18 + 0/–6	114 ± 13 ka	estimated max. depth of marine embayment	poor to moderate
–28.78 + 0/–6	(99 ± 11)	estimated max. depth of marine embayment	poor to moderate
	111 ± 12 ka		
lower than –37.97	119 ± 13 ka	loess soil and fluvial deposits	directional information only
–39.39 + 2/–0	136 ± 10 ka	barrier blocked swamp	very high
–45.75 + 2/–0	158 ± 19 ka (est.)	barrier blocked swamp	very high
higher than –51.12	180 ± 31	marine sediments	directional information only
higher than –52.37	172 ± 33	marine sediments	directional information only
lower than –61.12	204 ± 37	fluvial sediments	directional information only

Note that sea-level with an interpolated age in this table is not plotted in Fig. 9.

sea level must have been higher than the top of marine units such as Unit C, others are precise. These include values controlled by estuarine deposits, and by beach and back swamp deposits such as those of Unit D, which must have been very close to contemporaneous sea level. The presence of marine dinoflagellates suggests an open sea environment at the base of Unit C, while floral, and sedimentological characteristics suggest an interglaciation. The two TL ages fall within the MIS 7 interglacial at 180 ka. Within the postulated Stage 5 units, sea levels are derived from several lines of evidence, and are consistently higher than any levels that can be derived from younger sediments until the Holocene section of the curve is reached.

Throughout Stages 4 and 3 the floor of Gebbies Valley was dry land. Although several loesses are believed to have been deposited during the Otiran (Griffiths, 1973) only one of the appropriate age, Unit L, has been identified in the core. If others were deposited, they have presumably been eroded away, together with an unknown amount of the marine sediments of Unit K. Marine influences reappear with the onset of the Holocene transgression. Sea level is believed to have stabilised at about its present level around 6 ka (Gibb, 1986), and thereafter changes in sedimentation in Gebbies Valley are explicable in terms of changes of the coast outside the valley.

Using all the above lines of sea-level information, we have plotted the depth–age relationships and extrapolated the data onto a standardised oxygen isotope curve widely used as a proxy for sea levels for the last 150 ka (Shackleton, 1987; and confirmed recently by Chappell et al., 1996 (Fig. 9)). Our data show good agreement with the curve, with a strong clustering of ages occurring in Isotope Stage 5.

9. Repetitive pattern of coastal evolution as indicated by systematic changes in diatom assemblages

In both the early to mid Holocene and latter stages of the last interglaciation (MIS 5a), extensive diatom records are preserved. The systematic changes in the diatom assemblages are remarkably consistent between the two interglaciations. In both instances, the diatom bearing sediments are underlain by a marine transgression. In the case of the Holocene, this is marked by the appearance of aqueously deposited silts and clays over a loess, while the latest last interglacial sequence is underlain by a sub-littoral muddy unit believed to represent a marine transgression at the termination of MIS 5b.

In both instances, the initial diatom assemblages are largely composed of marine and brackish types.

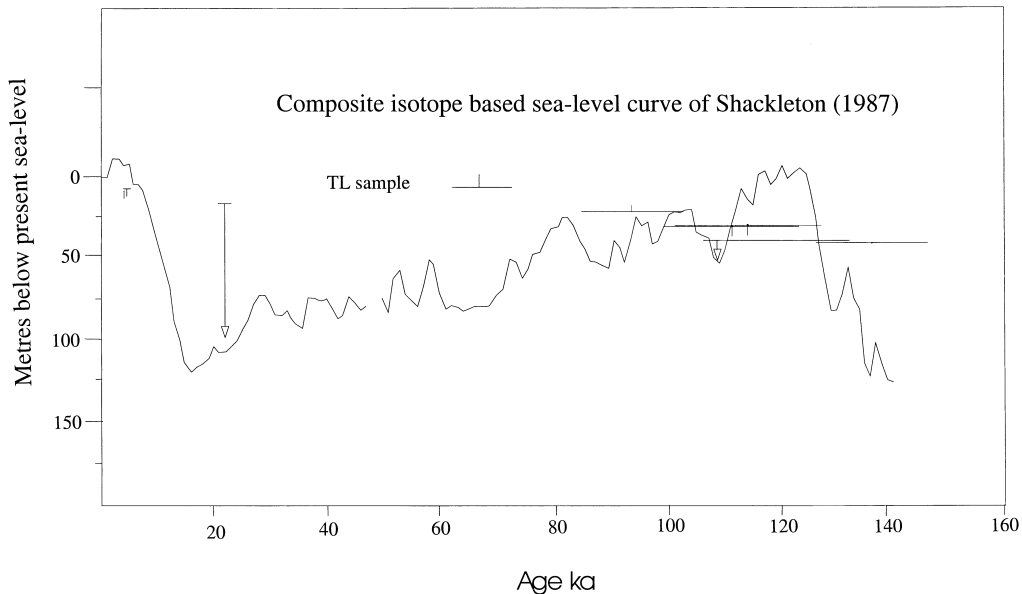


Fig. 9. Sea-level data from this paper plotted against the standardised sea-level curve of Shackleton (1987). Only securely dated levels from MIS 1–6 are plotted. Within the limitations of the age control there is a very good correspondence between our sea-level points and the curve, especially in Stage 5.

During the Holocene, these assemblages are dominated by *Nitzschia granulata* and *Diploneis subovalis/smithii*, while during Isotope Stage 5a, *Melosira westii* dominated (see Figs. 4 and 5). In both, the marine–brackish assemblage is replaced abruptly by a freshwater assemblage dominated by shallow water colonising species, particularly the small *Fragilaria* species and *Fragilaria brevistriata*. These types dominated for several metres with rare short-lived reversions to marine assemblages. These marine assemblages probably represent the aftermath of major river floods where a temporary breach in the barrier allowed significant post-flood incursion of saline waters. The initial freshwater assemblages are subsequently replaced by another freshwater assemblage composed of planktonic freshwater diatoms including various species of *Cyclotella* and *Aulacoseira*. More subtly, within the *Fragilaria*-dominated zone, there is a gradual increase in plant-attached diatoms at the expense of the colonising species. The close correspondence in the trends between the Holocene and MIS 5 strongly suggests that the diatom communities are responding to a repetitive form of environmental forcing in both interglaciations.

10. Barrier formation and lake development in the Ellesmere Basin during the Holocene

Several papers have been written about the evolution of the Canterbury coastline during the Holocene (e.g. Armon, 1973, 1974; Shulmeister and Kirk, 1993, 1997; Kirk, 1994). Most recently, Soons et al. (1997) have revised the Holocene history of the Ellesmere Basin and proposed a model of dynamic coastal change during the Holocene. Marine transgression during the terminal Pleistocene and early Holocene created a large embayment on the southern side of Banks Peninsula which was gradually cut off by the development of a spit that extended eastward from near the mouth of the Rakaia River. By some time in the mid-Holocene this spit cut off the embayment creating first an estuary and eventually a lake basin (now Lake Ellesmere). The ridges that formed the spit were close to sea level, probably not more than 1–2 m AMSL and the lake that formed behind them must initially have been very shallow.

Soons et al. (1997) have demonstrated that the closure was not permanent and that, at least intermittently, the lake was converted to a lagoon by the breaching of the barrier. Nevertheless, after the initial

formation of a spit ridge in the mid-Holocene, any further material transported along the coast would be used not to extend the ridge but to orientate the ridge into the swash direction. This was accompanied by an increase in ridge heights towards the modern beach, to at least +11 m AMSL.

The Holocene diatom record from Gebbies Valley is consistent with the inferred history outlined by Soons et al. (1997). The following biogeomorphic model of evolution is therefore proposed.

(1) During the early Holocene marine transgression, diatom flora in the Ellesmere embayment were composed of marine and brackish taxa with background levels of freshwater types washed into the embayment.

(2) The sharp termination of the marine assemblages is consistent with the rapid conversion of a salt water body to a freshwater body after the spit closed off the embayment mouth.

(3) In the immediate aftermath of the closure and conversion to freshwater, a bloom of freshwater colonising species occurred. These remained dominant for a prolonged period, though gradually, weed-attached species rose in prominence.

(4) In the late Holocene a further change occurred with freshwater planktonic types becoming dominant. These are important in deeper and/or more turbid waters. This is consistent with the increase in height of the barrier and a consequent rise in lake levels.

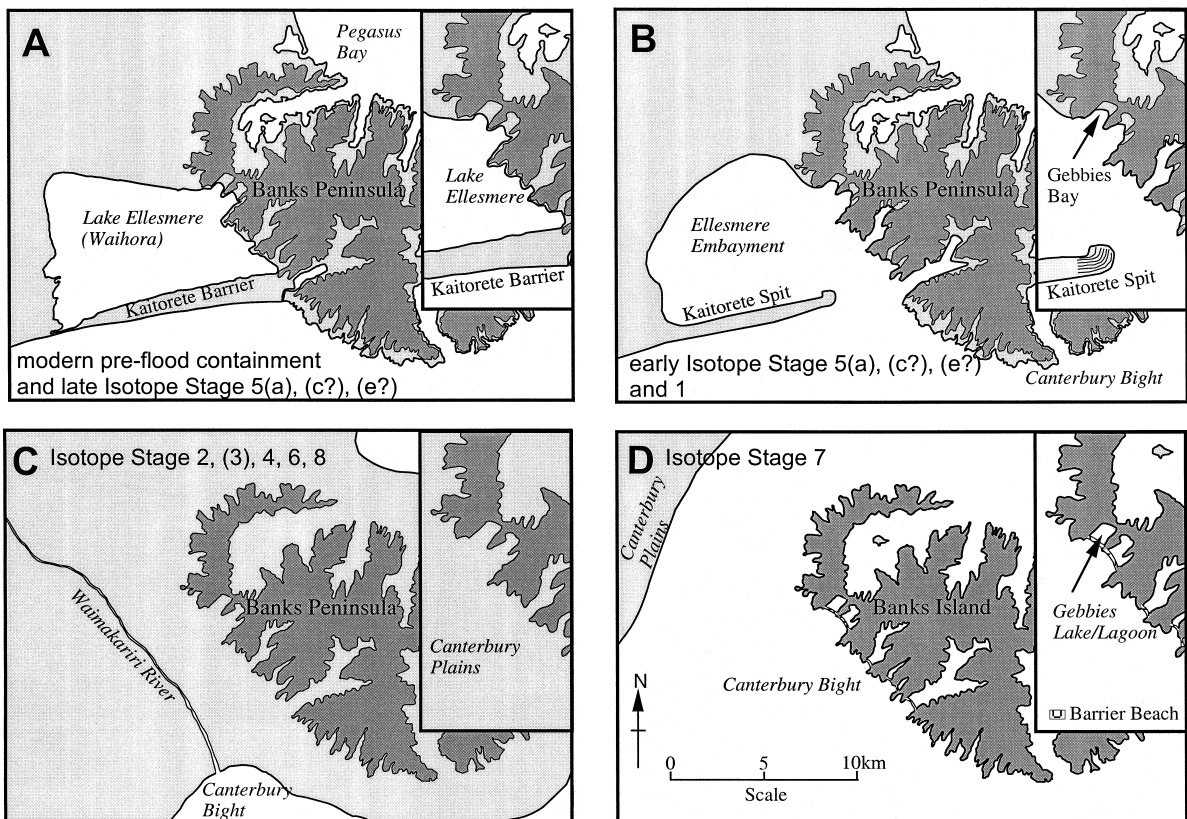


Fig. 10. Coastal evolution of the Kaitorete Spit during the last 200 ka. Inset map shows the area in the immediate vicinity of Gebbies Valley. (A) High sea-stand conditions late in transgressive (warm) periods with a barrier beach blocking off Lake Ellesmere. (B) High sea-stand conditions during transgressions and soon after high sea-stand has been achieved, with a spit growing out from the South Canterbury coast gradually closing off the Ellesmere Embayment. (C) Demonstration of low sea-stand conditions during stadials. The suggested coastline represents glacial maximum conditions. (D) Prior to Stage 5, Banks Peninsula was probably separated from the plains by a marine channel during interglacials. A small locally sourced barrier beach occurred in Gebbies Valley. Coastlines away from Gebbies Valley are schematic only.

The similarity of the Holocene diatom record with that from the close of the last interglaciation (MIS 5a) implies a similar barrier–lagoon development during that transgression. At the same time, the coastal deposits associated with Units D and C and attributed to MIS 6 and 7, respectively, suggest either a barrier beach inside Gebbies Valley, or no barrier at all. The implication is that the Kaitorete Spit did not exist prior to MIS 5. The Holocene barrier is fed from gravel quarried by wave action, from the alluvial fans of the Canterbury Plains, and its absence from palaeo-Lake Ellesmere suggests that these fans had not prograded far enough into the Canterbury Bight for gravel to be funnelled onto Banks Peninsula. Thus the final attachment of Banks Peninsula to the Canterbury Plains probably occurred during the low sea level of MIS 6. A summary planform diagram of regional palaeogeographies near Gebbies Valley and Lake Ellesmere through the last ca. 200 ka is displayed in Fig. 10.

11. Conclusions

The flanking valleys of Banks Peninsula contain substantial late Quaternary fills. These fills relate in the main to marine transgressions. Although there are large error estimates attached to our luminescence chronology, we have established that most of the sediments preserved are related to interglaciations, with less complete coverage during the intervening glaciations. The core provides a useful palaeoecological record from an under-recorded area, a detailed history of barrier and lagoon development along this coast over the last 200 ka and a MIS 5 sea-level record from the New Zealand region.

The vegetation history of the site indicates that past interglacial vegetation was very similar to the pre-European modern vegetation of the peninsula. By contrast, interstadial conditions along this coast are marked by a low forest–tall shrubland composed of mixed montane and coastal affinity plants. Elevated levels of charcoal and high values of *Leptospermum* pollen (a pioneer species of disturbance sites) in the interstadial deposits suggest that the plant communities were unstable, with fire being an important element. We tentatively attribute increased burning to increased wind stress under enhanced

foehn winds from the northwest. Our pollen record is not easily comparable to the marine record from DSDP Site 594. This may simply reflect contrasts between dominantly locally derived pollen at our study site and fluvially transported material at DSDP Site 594, or it may represent a real discrepancy.

Our results indicate that the barrier beach and lagoon–lake systems that exist on the southern side of Banks Peninsula during the Holocene recurred during previous high sea-stands, at least as far back as MIS 5. The Stage 5 pattern of evolution of this barrier–lagoon system is basically identical to the Holocene evolution of the system and this is interpreted to infer that Banks Peninsula was connected to the Canterbury Plains during MIS 5. By contrast, there is no conclusive evidence for the existence of the Kaitorete Barrier prior to MIS 5 and we conclude that Banks Peninsula was probably an island during earlier interglaciations.

Our summary sea-level data is generally consistent with the mounting evidence for the tectonic stability of Banks Peninsula during the late Quaternary. Combined with the known record of low sea-stands on the Canterbury and Otago shelves (e.g. Carter et al., 1986), this site provides an important late Quaternary sea-level record for this area of the southwest Pacific.

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