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A New Zealand climatic template back to c. 3.9 Ma: ODP Site 1119, Canterbury Bight, south-west Pacific Ocean, and its relationship to onland successions

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Abstract Ocean Drilling Program Site 1119 is located east of South Island, New Zealand, in 394 m water depth on the upper continental slope. Though 93 km offshore today, during recent glaciations the site lay near the lowstand shoreline, directly beneath the path of seasonal riverine meltwater plumes. The core comprises an upper portion (0–86.19 metres composite depth (mcd)) deposited as upper slope clinoforms and a lower portion (to 513.5 mcd) deposited as mid-slope, intermediate water depth sediment drifts, the two intervals being separated across a short c. 25 k.y. long unconformity within Marine Isotope Stage (MIS) 8. Almost throughout, the core comprises alternating micaceous muds and silts (glacials) and muddy, sometimes calcareous, sands (interglacials). Because of the enhanced potassium content of the terrigenous muds, the lithologies that characterise the two different climatic states possess markedly different natural gamma radiation signals, high and low, respectively. The natural gamma profile of Site 1119 therefore provides a high quality climatic time series with a 1–2 k.y. resolution (as sampled) back to the mid-Pliocene. The cyclic gamma ray pattern provides a proxy measure of ice volume in the Southern Alps, and is therefore an atmospheric record. This record matches closely that of oceanic oxygen isotope curves back to MIS 100 (2.4 Ma) and less closely beyond to MIS Gi-11 (3.91 Ma) at the base of the core. Marked, high gamma ray intervals at 3.68–3.63, 3.38, 3.12, and 2.80–2.67 Ma may reflect sharp mid-latitude atmospheric coolings at these times, as supported by marine faunal and isotopic evidence elsewhere in the New Zealand region. Alternatively, they may in part reflect changes in clay provenance consequent upon tectonic uplift in the hinterland. The natural gamma measurements are consistent with an overall decline of 6°C in average atmospheric temperature over South Island, New Zealand, since 2.45 Ma. Milankovitch 41 k.y. cyclicity is also prominent in the natural gamma record over this period, and its 60 API unit magnitude implies temperature swings up to 12°C between glacials and interglacials (G/I). Similarly large G/I temperature changes occur also in nearby sea surface temperatures east of South Island and in atmospheric temperatures in Antarctica (Vostok), suggesting that Pleistocene climate signals were closely synchronised across wide areas of Southern Hemisphere middle and high latitudes. Together with changing tectonic and related palaeogeographic patterns, these temperature changes must have played a significant role in influencing the local time ranges of taxa used in biostratigraphy. Very few, if any, traditional New Zealand Pliocene-Pleistocene index fossils have synchronous ranges across the region, and other methods of correlation (magnetostratigraphy, cyclostratigraphy, tephrochronology, isotope stratigraphy, and numeric dating)

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have therefore come to be of particular importance. The Site 1119 natural gamma record provides a unique reference template for climate variation in New Zealand mid-latitudes, with which are compared the available correlation markers for both onshore and offshore Pliocene-Pleistocene strata in the New Zealand region.

Keywords Ocean Drilling Program; ODP Site 1119; Canterbury Drifts; Canterbury Bight; climate change; natural gamma radiation; Pliocene-Pleistocene; stage; New Zealand

INTRODUCTION

This paper describes the cyclic natural gamma ray climate signature of ODP Site 1119, located 93 km offshore from Timaru, South Island, New Zealand, at a water depth of 394 m (Fig. 1). The gamma record is continuous back to c. 3.91 Ma in the middle Pliocene (Carter & Gammon 2004), apart from a c. 25 k.y. long unconformity within Marine Isotope Stage (MIS) 8, and small diastems several thousand years long at the transition between some glacials and interglacials (notably the boundaries of MIS 2/1, 6/5, and perhaps 12/11; Carter et al. 2004b). In attempting to summarise New Zealand Pliocene-Pleistocene climatic history, the stratigraphy of Site 1119 serves as a useful template against which to assemble, and with which to compare, other less complete successions.

An important New Zealand subantarctic water record which overlaps with that of Site 1119 is the nearby but deeper water Deep Sea Drilling Project (DSDP) Site 594, the continuous MIS 1-27 upper part of which may be truncated below by an unconformity (Nelson et al. 1985, 1993; see also core MD 97-2120; Pahnke et al. 2003). Further north, ODP Sites 1123, 1124, and 1125 (Carter et al. 1999) record the climatic history of cool subtropical waters. Onland marine records include the mainly interglacial strata of the Wanganui Basin (Saul et al. 1999) and the mostly deeper water sediments of the Mangaopari Basin (Kennett et al. 1971; Gammon 1997), both of which record a largely cool subtropical water history, although with subantarctic punctuations (cf. Fleming 1944; Vella & Nicol 1970) which more recent writers attribute to northward cold water current jetting (Nelson et al. 2000). Finally, terrestrial glacial deposits are widespread along both sides of the Southern Alps. Inevitably, however, these strata are spasmodically preserved, fragmentary, and heavily biased towards the record of the last few glaciations only (e.g., Suggate 1965; cf. Gibbons et al. 1984; Denton et al. 1999).

This paper is based on the presumption that, back to at least 4 Ma, and perhaps beyond, a local stage classification is no longer the most useful framework within which to pursue studies of the New Zealand Pliocene-Pleistocene. Nonetheless, because of the copious published information that is inherent in historic stage terms, these stages are shown on the scale of relevant figures. For reasons that relate to Historical background (see below), and are detailed in Carter (in press), the stage usages followed are those of Carter & Naish (1998) and not Beu (2001) or Cooper (2004).

HISTORICAL BACKGROUND

In the early 19th century, amid vigorous debate, an understanding developed that large parts of northern Europe had been buried under active ice sheets in recent geological times (Agassiz 1840). On emigrating to the United States of America in 1846, Agassiz, together with other early American geologists, showed that a similar conclusion held also for North America. Meanwhile, New Zealand's pioneer geologists were well in touch with this contemporary Northern Hemisphere thinking. Noting the presence around the South Island mountain fringes of rounded topography, deeply incised rock basin lakes and surficial glacial "drift" deposits (Fig. 2), they concluded that the Southern Hemisphere too had been subjected to episodes of expanded glaciation in the recent past (Haast 1864; Hutton 1872, 1876; McKay 1881).

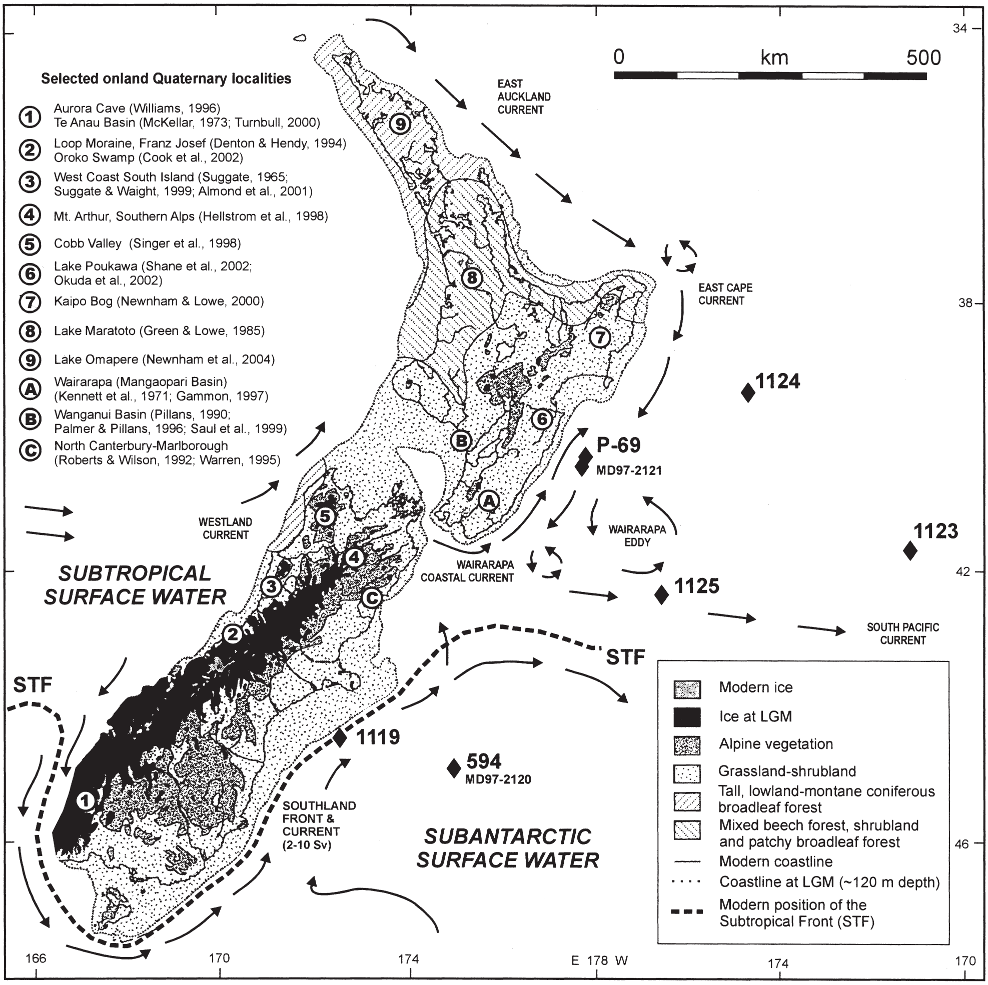

Fig. 1 Reconstruction of New Zealand geography at the peak of Marine Isotope Stage 2 (last glacial maximum), after Fleming (1975), McKinnon (1997), and Newnham et al. (1999). The location of selected onland Quaternary localities is indicated by letters and numbers (keyed, top left). Features indicated offshore include modern surface currents (arrows), the position of the Subtropical Front (heavy dashed line), and the locations of notable core sites (diamonds).

By the end of the century, the concept of a relatively recent "great glacier period" was firmly established in New Zealand (Hutton 1900), although Hutton himself believed that the largest glaciations must have preceded the deposition of the Pleistocene temperate water shellbeds and blue clays at Wanganui, and, therefore, occurred in the Late Pliocene (e.g., Hutton 1872).

Detailed mapping of glacial deposits in the European Alps continued throughout the late 19th century, culminating in the publication of the classic monograph by Penck & Bruckner (1909). These authors described the evidence for the passage of four major Quaternary climatic episodes (the Wurm, Riss, Mindel, and Gunz glaciations), each of which was linked to the deposition of successively higher (older) moraines and outwash terraces in mountain tributaries of the River Danube. Thereafter, and right up until the 1970s (e.g., Fleming 1976), this four-fold European stratigraphic framework exercised a pervasive influence on all studies of Quaternary sediments worldwide.

After publication of Penck & Bruckner's work, New Zealand Quaternary studies developed in two disparate ways. One strand of research concentrated on the terrestrial valley moraines and outwash trains of the Southern Alps and, pursuing a close correspondence with European thinking, interpreted the stratigraphy in terms of four major climate cycles, which came to be termed the Otiran, Waimean, Waimungan, and Ross glacial stages, or glaciations (Gage & Suggate 1958; Suggate 1965). Individual ice advances within each such stage carried their own named designation, as for the Kumara-2 to Kumara-4 advances that were recognised within the Otiran (e.g., Suggate 1990). At the same time, a second strand of Quaternary study became centred around the outstanding New Zealand record of uplifted Pliocene-Pleistocene marine strata. The abundant marine fossils (Marshall & Murdoch 1920; Powell 1931, 1934; Laws 1940; Fleming 1944, 1953; Beu 1965, 1995) throughout these and other older New Zealand Cenozoic sediments caused attention to focus on the development of a set of local "time" subdivisions, using the mostly endemic biostratigraphy to recognise a series of marine stages (Thompson 1916; Allan 1933; Finlay & Marwick 1940, 1947). Thus were introduced, in some cases with later substage divisions (Fleming 1947, 1953), the Opoitian (Finlay 1939), Waitotaran (Thomson 1916), Nukumaruan (Marwick 1924), Castlecliffian (Thomson 1916), and Haweran (Thomson 1917; Beu et al. 1986) Stages, which together encompass the middle Pliocene-Pleistocene-Holocene stratigraphic interval as it is recognised today. As defined and used in New Zealand, these units were actually biostratigraphic oppelzones (multi-taxon, concurrent-range zones) (Carter 1974), but nonetheless use of the term "stage" has persisted.

By the 1970s, New Zealand stratigraphers had become accustomed to their binary scheme of marine/non-marine Quaternary subdivision, and new techniques of correlation that developed at the time, such as magnetostratigraphy, tephrostratigraphy, numeric isotope dating, and amino acid racemisation, were comfortably absorbed within it. Meanwhile, however, international study of the Quaternary had moved on, stimulated by the demonstration by Emiliani (1955, 1966) and Shackleton & Opdyke (1973, 1976) that the universally used "four glaciation" classification was a chimera. These authors showed that, instead, about 14 major glacial/interglacial (G/I) cycles had occurred back to c. 1 m.y., a record that ODP drilling later showed to continue downwards to comprise in total 50 G/I cycles back to c. 2.5 Ma (e.g., Raymo et al. 1989; Ruddiman et al. 1989). The rhythmic regularity of these climatic oscillations was immediately apparent, and was shown by Hays et al. (1976) to correspond to the 20, 41, and 100 k.y. long Milankovitch periodicities that are associated with Earth's orbital changes. Quickly, Kukla (1977) provided the first convincing integration between the classic European glacial climatic history and the marine oxygen isotope record, and finally Gibbons et al. (1984) furnished an elegant explanation of the enigma, given the passage of at least 50 major G/I cycles since 2.5 Ma, that continental deposits worldwide seem to preserve a record of only four or five major glaciations. Meanwhile, and whilst many conventional Quaternary stratigraphers continued to squabble about the best position in which to define the Pliocene-Pleistocene boundary (laid to rest in Aguirre & Pasini 1985), active research practitioners in fields as disparate as testing the accuracy of marine microfossil zonation (Thierstein et al. 1977) and hominid palaeontology and cultural development (Stringer & Grun 1991) simply adopted the astronomically-tuned marine isotope stage record (Shackleton et al. 1990; Tiedemann et al. 1994) as their de facto time framework for Quaternary studies.

The application of these developments to New Zealand stratigraphy started with the insightful review paper of Beu & Edwards (1984), and has continued apace ever since. Therefore, as pointed out by Suggate & Waight (1999, p. 78) with respect to the New Zealand glacial stages, local Quaternary time units "are rapidly becoming redundant with the increasing use of

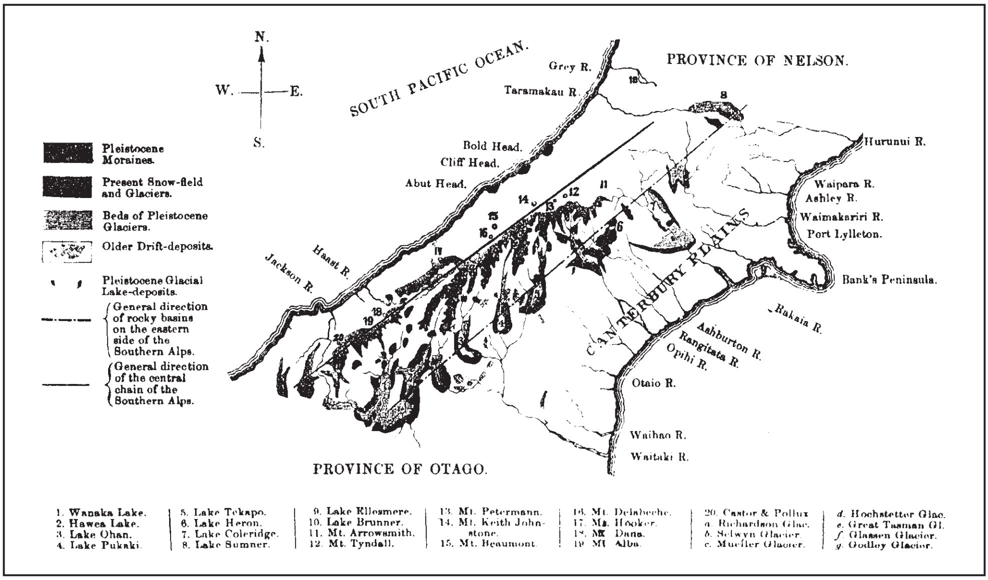

Fig. 2 Sketch map of the Province of Canterbury, "showing the Glaciation during Pleistocene and Recent times, as far as explored" (Haast 1864, p. 134). The two darker shadings are indistinct on all versions of this historic figure and its key. Note, therefore, that (i) the shading of the upper dark grey box ("Pleistocene Moraine") refers to the arcs of glacial debris located at the downstream end of the major valleys on both the east and west side of the Southern Alps; (ii) the shading of the lower dark grey box ("Present Snow-field and Glaciers") corresponds to the dark hachuring present along the mountain spine of the Southern Alps.

direct correlation of glaciations and interglacials with deep-sea oxygen isotope stages". This prophetic statement quickly became true. For instance, all non-marine Quaternary geology mapped on the revised QMAP 1:250 000 map sheets (e.g., Turnbull 2000; Nathan et al. 2002) is now presented in an MIS framework using the acronym OIS for oxygen isotope stage. And, to take but one marine example, Pillans et al. (1996) have provided a powerful demonstration of the accuracy of using isotope stratigraphy to date ashes in New Zealand marine cores compared with the many other numeric dating and correlation techniques that are available. Building on the discussions of Vella (1975) and Carter & Naish (1998), therefore, the time has arrived for the MIS terminology to be employed routinely in the analysis of New Zealand Quaternary successions, marine and non-marine alike.

ODP SITE 1119, EASTERN SOUTH ISLAND

ODP Site 1119 is situated 93 km offshore from eastern South Island at 44°45.3′S, 172°23.6′E. The site lies on the upper continental slope, about 5 km downslope from the shelf edge, and is located south of Banks Peninsula and almost due east from Timaru (Fig. 1). The stratigraphic setting and sedimentation history of the site are summarised in Fig. 3 and 4. Depths are referred to in terms of metres below the seafloor (mbsf, as measured during drilling; Carter et al. 1999, tables T1–T7), metres composite depth (mcd; an integrated common depth scale, prepared shipboard, which can be applied to all three of the cores drilled; Carter et al. 1999, individual site summaries), or revised metres composite depth (a slightly modified scale which smooths over four small gaps present between 122.74 and 169.55 on the mcd scale; Carter & Gammon 2004, supporting online material). Further site details and shipboard stratigraphic results are available in Carter et al. (1999).

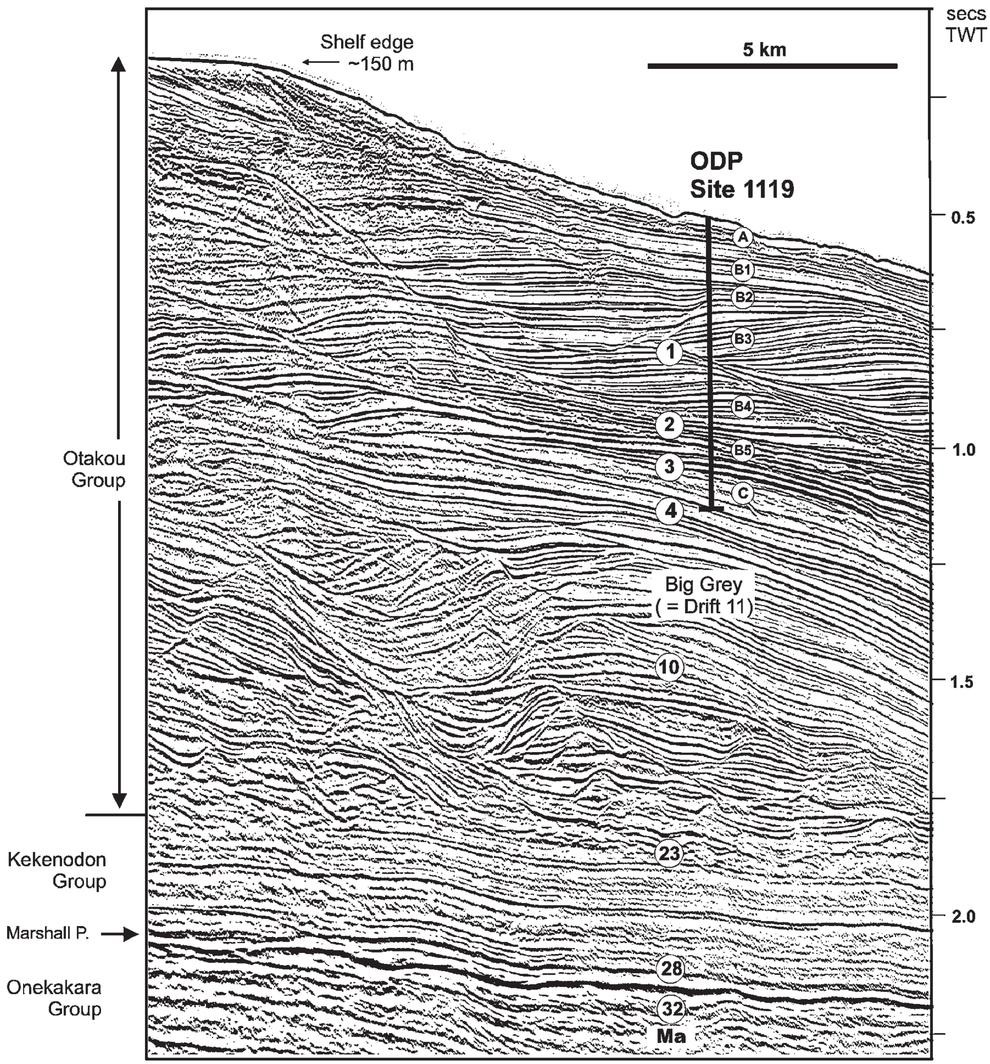

Fig. 3 Seismic section south-east through ODP Site 1119. Part of high resolution multichannel profile EW00-01 (R/V *Maurice Ewing* cruise; Lu & Fulthorpe 2004). Three distinctive intervals in the upper part of the Otakou Group were penetrated during drilling, as labelled to the right of the drillsite: **A**, upper slope clinoforms, which represent progradation of shelf-connected deltaic foresets during glacial lowstands, with intercalated interglacial sand-mud veneer, deposited beneath STW, STW/SAW transition, and SAW waters; **B**, small mid-slope sediment drifts and drift tails, deposited obliquely along slope (upslope-climbing) from northward-travelling SAMW (upper part of interval: drifts) and transitional SAMW/AAIW (lower part of interval: drift tails); **C**, large lower-mid-slope sediment drifts, deposited along slope from northward-travelling AAIW. Approximate age (My) indicated by circled numbers to the left of the drillsite, continuing vertically below.

Site 1119 was located so as to provide information regarding southern mid-latitude climate history (waxing and waning of the Southern Alps ice cap; Carter & Gammon 2004), the relationships between sediment drift deposition, climate cycling and intermediate depth water flow strength (Carter et al. 2004a), and ocean front and watermass migrations (Carter et al.

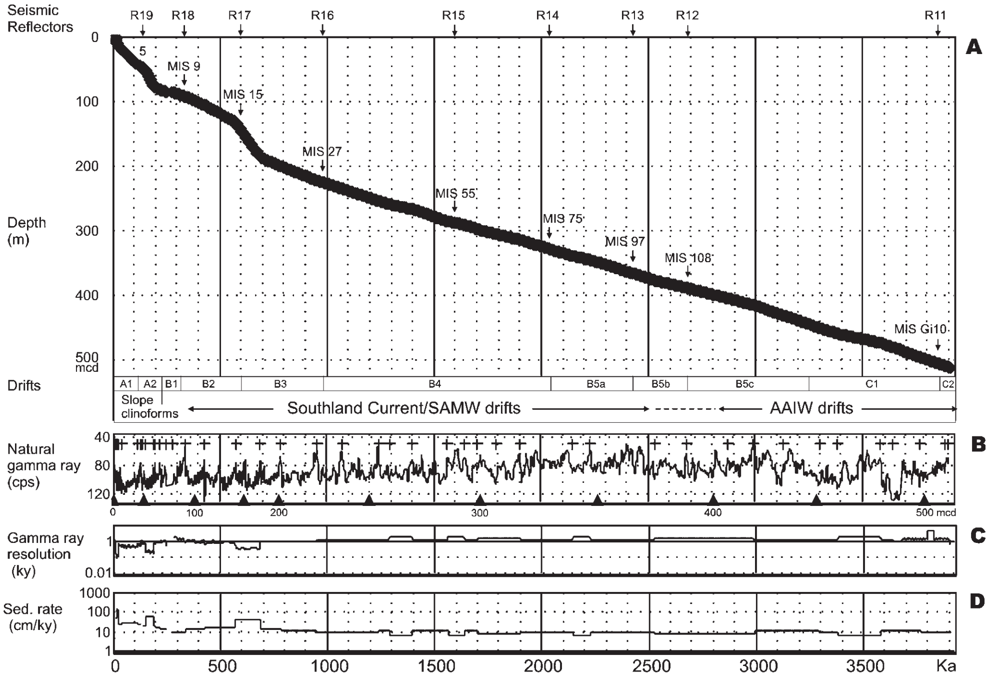

Fig. 4 **A**, Time-depth diagram for ODP Site 1119, with positions of seismic reflectors (top), selected oxygen isotope stages (labelled arrows, on main curve and sediment drifts (A1–C2) indicated; **B**, composite natural gamma ray curve with time-scale tie-points indicated by +, and 50 m increments in metres composite depth (mcd) denoted by black upward-pointing triangles (after Carter & Gammon 2004); **C**, resolution of the natural gamma curve; **D**, sedimentation rate in cm k.y.$^{-1}$. Both C and D are plotted on logarithmic scales. Age model after Carter & Gammon (2004). Reflectors R11–R19 after Lu & Fulthorpe (2004). Marine Isotope Stage (MIS) terminology after Shackleton et al. (1990, 1995).

2004b). Building on these understandings, the present paper uses the high resolution natural gamma record from Site 1119 as a dated template against which to consider other Pliocene-Pleistocene data from the New Zealand region. Though derived from the marine realm, the natural gamma curve from Site 1119 reflects atmospheric climate change, thus making the core uniquely suitable for studies that aim to integrate terrestrial and marine climatic events. Furthermore, unlike most land based records, such as that of the Wanganui Basin where interglacial sediments predominate and glacial strata are largely missing (Beu & Edwards 1984), the Site 1119 core preserves a nearly complete G/I stratigraphy.

Oceanographic setting

Marine waters off eastern South Island are dominated by northerly circulation and by the presence of the Subtropical Front (STF), which separates cold subantarctic water (SAW) offshore from warmer cool subtropical water (STW) on the shelf (Garner & Ridgway 1965; Heath 1972, 1981, 1985). In the open ocean on either side of New Zealand, the STF runs along the Chatham Rise at latitude 43°S and across the southern Tasman Sea at latitude 46°S. These two east-west frontal segments are connected along the eastern side of the South Island by the north-easterly oriented Southland Front (Chiswell 1994).

The Southland Front runs parallel to the bathymetry about 80 km offshore and just seawards of the shelf edge, above seabed depths of 200–400 m (Fig. 1). Inboard of the front, the shelf watermass is driven northwards by prevailing southerly storms and by the north flowing inshore part of the Southland Current (Jillett 1969; Chiswell 1996) which carries >34.5‰ salinity STW sourced from the East Australian Current into the Pacific Ocean. The inshore Southland Current entrains subantarctic water through mixing across the Southland Front (Heath 1972). Recent work (Sutton 2003) has shown (i) that the core of the current lies seawards of the Southland Front and within SAW; (ii) that the current attains a mean northward velocity greater than 20 cm s^{-1}, with 10 cm s^{-1}motions reaching down to 750 m depth; and (iii) that the mean transport of 8.3 Sv comprises about 90% SAW and only 10% STW. The Southland Current is, therefore, the probable emplacement agent for the shallower parts of the Canterbury Drifts described by Carter et al. (2004a,b). Beneath the surface watermasses, cold, intermediate depth waters also travel northward along the eastern New Zealand margin (Morris et al. 2001). Subantarctic Mode Water (SAMW), formed by seasonal convection at the Subantarctic Front, covers the entire Campbell Plateau and is present also at slope depths between c. 250 and 600 m, moving within the Southland Current. SAMW grades down into Antarctic Intermediate Water (AAIW), which lies between c. 800 and 1100 m and is derived by subduction at the Antarctic (or Polar) Front. These watermasses and their motions exert a controlling influence on sediment transport and deposition off eastern South Island.

Sedimentary setting

Sedimentary material is shed southward or eastward from the Southern Alps into eight major river systems, four of which coalesce to form the 300 km long braid-plain of the Canterbury Plains. These rivers deliver c. 12 Mt of terrigenous sediment annually to the east coast shoreline (Griffiths & Glasby 1985; Hicks & Shankar 2003). On entering the marine environment, the sediment is transported northward. At shallow depths, northward motion is in response to the Southland Current and to the dominant southerly weather patterns which drive shelf surface waters (Carter & Herzer 1979); and sediment which is bypassed to slope depths between c. 300 and 1100 m encounters north-flowing geostrophic SAMW and AAIW (Morris et al. 2001). This interaction between the western sediment source and the dominant northward marine water motion forms part of a larger sedimentary system, termed the Eastern New Zealand Oceanic Sedimentary System (ENZOSS) by Carter et al. (1996). Off eastern South Island today, this system comprises four shelf-parallel sedimentary provinces (Carter et al. 1985). These comprise (i) at the coast and along the inner to middle shelf, a shore connected prism of terrigenous sand and embayment mud (Andrews 1973; Herzer 1981); (ii) on the outer shelf, and spilling onto the upper slope, a terrigenous-sediment-starved belt of carbonate sand and gravel, sourced mainly from benthic biota (Orpin et al. 1998), within which shelf parallel sand ribbons and north-facing megaripples testify to the presence of episodic strong currents (Carter et al. 1985); (iii) on the upper slope, a clinoform drape of mud derived from sediment bypassing over the shelf edge (Browne & Naish 2003; Carter et al. 2004b); and (iv) on the middle slope, in waters of intermediate depth, a major locus of sediment drift deposition which comprises alternating sand and mud of the Canterbury Drifts (Fulthorpe & Carter 1991; Lu et al. 2003; Carter et al. 2004a,b; Lu & Fulthorpe 2004).

Today, Site 1119 is located just downslope of the c. 150 m deep shelf edge, beneath the core of the Southland Current and near the top of the mid-slope belt of SAMW sediment drifts (Fig. 3), though modern SAMW flow appears subdued. The abundant terrigenous mud at the site is derived by unmixing from the outer shelf seabed during storms and bypassing over the shelf edge, where it accumulates as an upper slope mud drape at depths below storm wave base and above the main mid-lower slope SAMW–AAIW flow.

Effect of climatic oscillations

Early piston cores from the outer shelf and upper slope of eastern South Island demonstrated a marked difference between a surficial veneer of brown or olive oxygenated and often sandy and shelly Holocene sediment, and underlying blue grey, reduced, micaceous muds of latest Pleistocene age. A similar pattern also occurs back through the older climatic cycles penetrated in longer cores, where biopelagic carbonate proves to be characteristic of warm climate intervals, and radiolarians+mica characterise cold intervals (Griggs et al. 1983). This bipartite lithological pattern was dramatically confirmed and extended with the drilling of DSDP Site 594 on the Canterbury middle slope. Results from this site (Nelson et al. 1985) showed that the alternating carbonate/terrigenous couplets stretched back to the Brunhes/Matuyama boundary and beyond, and that isotope stages back to MIS 30 were well delineated by both their isotope and carbonate signatures. For the first time, the climatic cyclicity present in New Zealand Pleistocene sediments was shown to match closely with the global pattern established from other oceanic oxygen isotope records.

At recent sea-level lowstands, the Canterbury plains extended far offshore towards Site 1119 (Fig. 1), which was then located within c. 20 km of the shoreline and subject to the direct influence of glacial riverine sediment plumes. Not surprisingly, therefore, the stratigraphy of the site exhibits alternating terrigenous muds (glacial) with rapid sedimentation rates (>100 cm k.y.$^{-1}$ during parts of MIS 2) and muddy, calcareous sands (interglacial) (Carter et al. 2004b). This lithological cyclicity is similar to that described above from other nearby cores, and is ideally suited to characterisation by its natural gamma ray signature. In addition, the Site 1119 core throughout has a higher sedimentation rate (mostly between 10 and 30 cm k.y.$^{-1}$), and therefore better resolution than previous cores, and an almost unbroken record back to the middle Pliocene.

The natural gamma record from Site 1119

Data collection

Natural gamma records during Leg 181 were compiled both for the retrieved core, using the onboard multi-sensor track (MST) assembly, and *in situ* during downhole logging (DHL). Compiled shipboard data are available from Carter et al. (1999) and the composite gamma curve from Carter & Gammon (2004, supporting data). The onboard MST gamma counter uses four sodium iodide scintillation counters. The plotted MST output from Site 1119 comprises an 89.4% complete record (i.e., core loss was 10.6%) of total natural gamma variation in counts/s for the interval drilled, 0–494.8 mbsf (0–513.5 rmcd) (Carter et al. 1999, fig. F10). Downhole logging was performed between 85 and 492 rmcd with a Schlumberger natural gamma tool, which utilises a sodium iodide scintillation detector to measure the formation radiation and 5-window spectroscopy to resolve the detected spectrum into the three commonest naturally occurring radioactive elements: potassium (^{40}K), thorium (^{232}Th), and uranium (^{238}U) (Fig. 5A,B). Alternatively, the signal may be displayed as a total natural gamma ray count (Fig. 5C). Also plotted here (Fig. 5D,E) is the composite 0–513.5 rmcd natural gamma record assembled by Carter & Gammon (2004), which comprises a splice of MST/DHL/MST (0-107/107-487/487-513.5 rmcd) that covers the entire hole. This record, and its spectral composition, are discussed in more detail below. Note that all gamma records are plotted with reversed values on the y-axis, in order to facilitate comparison between the pattern that they exhibit and other standard climatic curves, such as the deep sea oxygen isotope record.

Significant differences in pattern occur between the MST and DHL gamma logs where they overlap (85–492 mcd) (Fig. 5C,D). These differences derive, first, from the fact that DHL measurements are derived from a greater volume of rock (c. 50 cm wall penetration) than

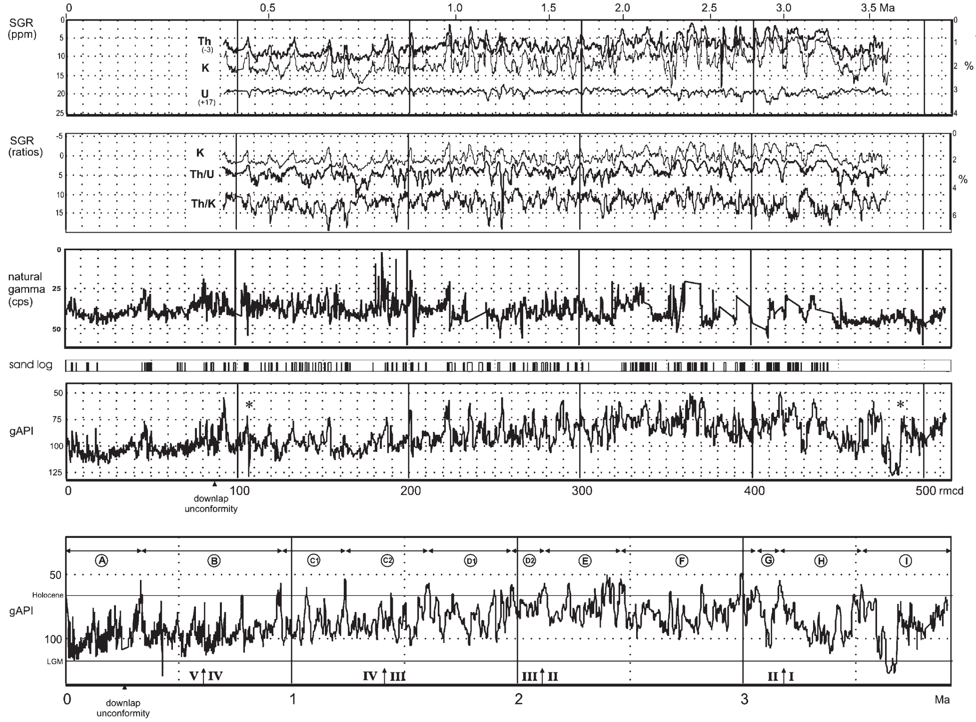

Fig. 5 **A**, Downhole spectral natural gamma ray measurements for Th, K (right-hand % scale) and U for Site 1119, plotted against depth between 91 and 479 rmcd. Curves separated for clarity by the addition of +17 for U and –3 for Th; **B**, Th/U and Th/K ratios calculated from data in A, plotted against depth, and compared with K content (right-hand % scale); **C**, total natural gamma ray counts/s for Site 1119 plotted against depth (rmcd). Measurements made on whole round core with the onboard multi-sensor track (MST) assembly, after temperature normalisation (after Carter et al. 1999); **D**, sand log for Site 1119, assembled from core, log, and Formation Microscanner (FMS) resistivity records. Sandy intervals with low gamma counts correspond mostly to interglacials; **E**, composite total natural gamma ray record (API gamma units), assembled by splicing MST sections top and bottom with the downhole log record; splice junctions indicated by two asterisks (after Carter & Gammon 2004); **F**, composite natural gamma curve displayed in E, but here plotted against age. Multi-isotope stage climatic cycles A–I labelled above the curve, and climatic episodes I–V of Marlow et al. (2000) labelled below the curve. The age model is based on AMS radiocarbon dates over the last 39 000 years (Carter et al. 2004b); below that level, 38 selected peaks have been mapped onto the 1119 gamma record from the tuned oxygen isotope records from ODP Sites 758 and 1143 (Carter & Gammon 2004). Note that the y-axis scale is reversed for all natural gamma curves (Fig. 4, 5) to display their similarities with other conventional climate change series such as oxygen isotope curves.

MST measurements (7 cm diam. core), and, second, from statistical variations that may be present in both records. Large variations between measurement runs may occur particularly in the DHL spectra, given their compilation from the high energy part of the spectrum which contains only c. 10% of the total count rate. Despite these differences, however, the MST and DHL records generally agree at the level of the major G/I spectral variations.

In interpreting gamma ray spectra, it is useful to remember that: K is most commonly associated with the presence of potash feldspars, micas, and illitic clays, but can also be produced by algal material or glauconite; high Th generally signifies the presence of heavy minerals such as monazite or zircon, but also occurs in some clays, in authigenic phosphates and in ash

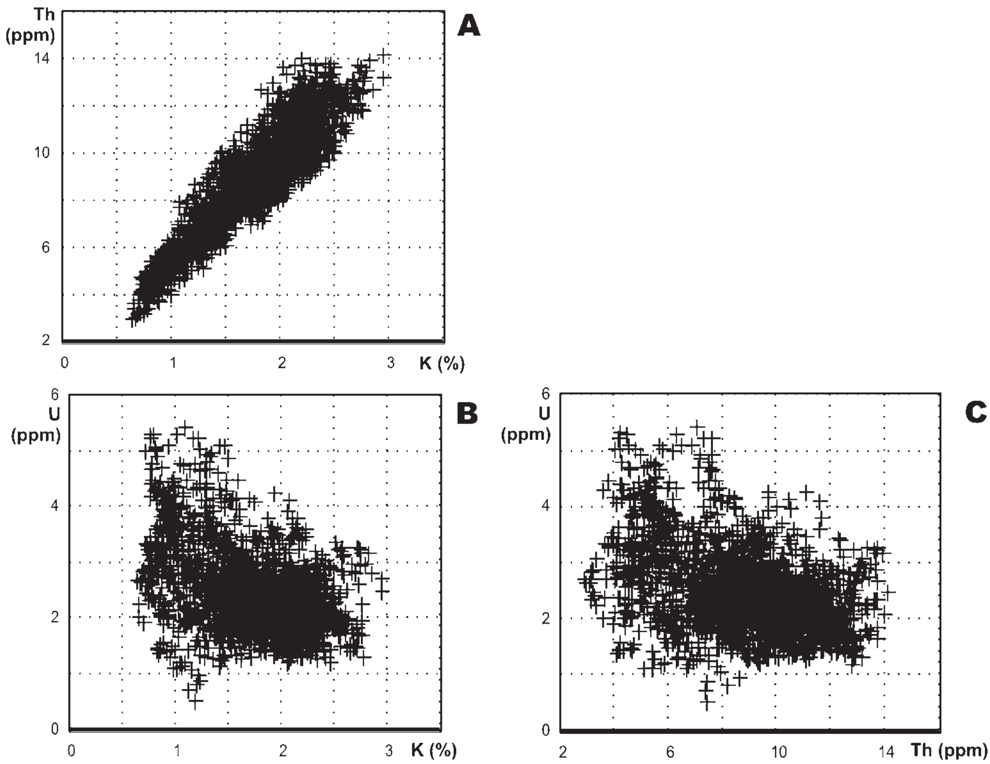

Fig. 6 Scatter plots based on downhole spectral natural gamma ray measurements at Site 1119 (Fig. 5A) for relationships between **A**, thorium (Th) and potassium (K); **B**, uranium (U) and potassium (K) and; **C**, uranium (U) and thorium (Th).

layers; and high U is often associated with high organic matter and/or condensed sediments, but also occurs within the heavy minerals suite. Because K and Th are the principal radioactive elements present in clays, other things being equal, a high overall gamma ray signature is associated with mud rich rocks, and a low gamma ray signal signifies the presence of sand or limestone.

Results

The summed natural gamma ray record mostly varies between about 50 and 110 API units, with extreme values of 42 API (2.41 Ma) and 128 API (3.65 Ma), and exhibits a marked cyclicity (Fig. 5E). Carter & Gammon (2004) have shown that the major peaks of this cyclicity occur at 100 and 41 k.y. Milankovitch periodicities, based on comparison between the gamma record and the coeval Vostok deuterium and marine oxygen isotope records. The spectral gamma record shows sympathetic fluctuations in K and Th against a relatively invariant U background (Fig. 5A). Scatter plots (Fig. 6) and the antithetic nature of the Th/U and Th/K curves (Fig. 5B) confirm a strong linear relationship between Th and K ($Th = (4.41*K) + 1.11$, $r^2 = 0.84$), and a lack of correlation between either of these elements and U. At the level of individual glacial-interglacial cycles, many parts of the gamma record show an asymmetrical saw-toothed shape similar to that of oxygen isotope records (cf. Broecker & van Donk 1970), with a sharp decrease in gamma at glacial/interglacial boundaries, followed by an irregularly stepped increase towards the peak of the ensuing glaciation. This pattern, which is generalised as a "funnel-shaped" parasequence (viewed uphole) in petroleum industry literature (Rider 1990), typically results at Site 1119 from the presence of sand in the interglacial intervals that coincide with the wide funnel tops.

Calibration of the Site 1119 gamma record against the Vostok deuterium (proxy Antarctic air temperature) record indicates that a 10 API change in gamma corresponds to an atmospheric temperature change of 2°C (Carter & Gammon 2004). The short term Milankovitch climatic swings at Site 1119 exhibit gamma changes of up to 60 API units, which is consistent with changes in air temperature between G/I as large as 12°C. This is approximately twice the traditionally accepted figure of 5–6°C atmospheric cooling during South Island glaciations, calculated from estimates of snowline depression (Willett 1950; cf. Porter 2001) or pollen analysis (Mildenhall 1995). Furthermore, and though it is a tropical location, Smith et al. (2005) have recently estimated a Last Glacial Maximum (LGM) temperature depression of only 2–3°C from studies of moraines in Peru and Bolivia. Nonetheless, the apparently large 12°C figure is consistent with several other independent temperature change estimates. The first is from the Vostok and Mt Fuji ice cores, where a 12°C Antarctic atmospheric temperature difference is inferred between, for example, interglacial MIS 9 and glacial MIS 8 (Watanabe et al. 2003). Second, planktic foraminiferal census determinations from Sites 594 and 1119 indicate G/I SST differences back to 0.9 Ma of up to 9°C at Site 1119 and 14°C for Site 594 (Wilson et al. 2005). Third, G/I temperature differences back to 0.35 Ma of c. 9°C are independently confirmed by Mg/Ca measurements from core MD 97-2120 (a reoccupation of the upper part of Site 594) by Pahnke et al. (2003). Fourth, and last, Beu (1999) has recently described a large northward extension of the known LGM range for the cold water scallop *Zygochlamys delicatula*, based on dredged samples from off Northland. Though Beu prefers to attribute the cold water needed to explain this occurrence to LGM upwelling, he points out that "conceivably…the new records of *Z. delicatula*…imply that glacial temperature minima fell by 1–7°C more than has been suggested previously", i.e., by up to 12°C.

In addition to the Milankovitch-scale cyclicity, the composite total gamma record (Fig. 5E) also exhibits longer term trends. Comparison of the individual spectral gamma curves shows that these trends result from changes in K and Th, and not U, content. First, above 370 rmcd (MIS 97; 2.45 Ma), the SGR curve exhibits a long-term trend of increasing mean gamma by about 30 API units, consistent with a long-term mid-latitude atmospheric cooling of c. 6°C. This is similar to the overall c. 7°C oceanic SST decline over the same period indicated by alkenone data from ODP Site 1084 at latitude 25°S in the South Atlantic Ocean (Marlow et al. 2000). Second, superimposed on the long-term increase in gamma signal (and climatic decline) since 2.45 Ma, and also extending back to the 3.9 Ma base of the hole, are a number of shorter multi-cycles. Lettered A–I on Fig. 5F, these encompass groups of several to many oxygen isotope stages. They range in length from 200 to 600 k.y. and comprise broadly symmetric cooling-warming multi-cycles 30–40 API units (6–8°C temperature equivalent) in magnitude. Similar climatic multi-cycles are also reflected in the 1119 carbon isotope record as saw-toothed or symmetric patterns of $\delta^{13}C$ depletion (Carter & Gammon 2004). Over the upper part of the Site 1119 record, these carbon isotope cycles were interpreted by Carter et al. (2004b) as resulting from movements of the watermasses which flank the STF as the front changed its position in response to fluctuating climate and sea level. Multi-cycles F–I, below 2.45 Ma (370 rmcd; Fig. 5F), appear to indicate especially strong Late Pliocene cooling, with cool maxima of 50–60 API units magnitude at 2.72 Ma (392 rmcd; MIS 110), 3.20 Ma (430 rmcd; MIS 132), 3.38 Ma (460 rmcd; MIS 146), and 3.67 Ma (480 rmcd; MIS Gi-4). Whether these "coolings" were really of the magnitude implied by their NGR signature, and their significance, is discussed further below.

Though Site 1119 gamma multi-cycles A–I are almost certainly climate-related, the degree to which they correspond directly with atmospheric patterns, or alternatively may mark oceanographic changes such as frontal migration or cycles in sediment provenance, remains undetermined. More widely, however, similar multi-G/I climatic cycles have been described from the upwelling west African margin of the South Atlantic Ocean. Here, Marlow et al. (2000) use SST estimated from alkenone data to recognise four phases, I–IV, of successive cooling since 2.5 Ma. Two older phases (II and III) match within the probable error of the age models with gamma multi-cycles G–E and D from Site 1119, respectively, and the two youngest phases (IV and V) are separated across the MIS 14/13 climate shift at 0.5 Ma, which marks the peak cold reached during multi-cycle B (Fig. 5F).

Interestingly, the last 0.75 Ma ocean temperature record at site 1084 exhibits a 4–5°C SST warming trend, as also seen at Sites 594 and 1119 (Wilson et al. 2004), whereas the gamma record at 1119 continues its long-term decline over the same period (albeit with an included slight warming oscillation between 0.5 and 0.3 Ma). That mid-latitude ocean temperatures warmed whilst atmospheric temperatures apparently continued to cool over this period is consistent with the occurrence of major ocean circulation changes around 0.8–0.7 Ma, and raises interesting questions about changes also in heat coupling between the atmosphere and oceans. However, given that Sites 594–1119 and 1084 are located several thousand km apart on a major flow path of the global thermohaline circulation, the similarities in their temperature records are consistent with climatic and oceanographic linkages that were at least hemispheric in extent. Finally, the match between the natural gamma estimate of up to 12°C in atmospheric cooling for large G/I cycles and nearby SST (Wilson et al. 2004) and Antarctic atmospheric temperature changes (Watanabe et al. 2003) of similar magnitude builds confidence in the result, which contrasts sharply with an estimate of only 2–3° for LGM cooling in the tropics (Smith et al. 2005). It appears that over wide areas of the middle and high latitudes of the Southern Hemisphere, Pliocene-Pleistocene climate signals were closely synchronised in both timing (Carter & Gammon 2004) and pattern detail (Carter et al. 2004b), whereas the tropics may have danced to a different tune.

Provenance control on the NGR signal

Terrigenous sediment that reaches Site 1119 is of low grade metamorphic provenance and almost exclusively derives from central and eastern South Island (Carter et al. 1999). The onland basement geology comprises metagreywacke and chlorite-zone schist, with belts of higher grade biotite schists and oligoclase schists restricted to the axis of the Haast Schist zone of Central Otago and the recently uplifted Alpine Schist along the Alpine Fault, respectively (Suggate et al. 1978). Major element X-ray fluorescence analysis of glacial-interval muds from Site 1119 shows that they have relatively high K_2O values up to 3.5%, and clay contents up to 45% (unpubl. data). This, and the often micaceous nature of the muds (Griggs et al. 1983), are consistent with the gamma ray variations representing changes in the amount of detrital mica and clay present. A cyclic signature is apparent also in other log data from the site, with density and resistivity logs reflecting clearly the lower porosity of glacial clays compared with their sandy, interglacial counterparts (Carter et al. 2004a).

deMenocal et al. (1992) described cyclicity of Milankovitch periodicity in natural gamma records from ODP Site 798 in the Sea of Japan, where, however, the terrigenous content of the high gamma clay rich layers is interpreted to result from glacial drought, erosion of nearby continental loess, and aeolian transport. In contrast, and although glacial sediments at Site 1119 may contain an aeolian component, most of the high gamma muds were probably delivered almost directly to the site from seasonal meltwater in jetting lowstand river plumes (Carter et al. 2004b).

DISCUSSION

Late Pliocene climate in the New Zealand region

Prior to 2.45 Ma, multi-cycles F–I in the Site 1119 gamma data encompass a number of marked Late Pliocene "cooling" excursions (Fig. 5F). A similar though much more subtle modulation of global ice-volume or temperature may be discerned on some parallel oxygen isotope curves, but the gamma excursions are of far greater magnitude (Fig. 7, see insert). Mainly because of such differences, the pre-2.45 Ma segment of the 1119 gamma curve is less similar to oceanic isotope curves than is the part younger than 2.45 Ma (MIS 97), and alternative interpretations are possible. For instance, the major gamma high at around 3.65 Ma, interpreted as Gi-2-4 on the Carter & Gammon timescale, might instead represent the conspicuous Gi-16 (4.0 Ma) isotope enrichment seen at ODP Site 846 (Shackleton et al. 1995). The precise dating of the older portion of the 1119 core is, therefore, less certain, though its general Late Pliocene age is demonstrated by microfaunal evidence (Hayward in Carter et al. 1999). In this context, and accepting the timescale assigned to Site 1119 at face value, I now summarise briefly the independent evidence that exists for Late Pliocene cooling in the New Zealand region.

Evidence for Late Pliocene cooling in New Zealand

There is long-standing and strong evidence for several pulses of Late Pliocene cooling in the New Zealand region. This inference is based on five main lines of evidence.

Direct evidence for Late Pliocene onland glaciation is represented by fault-involved glacial till, varved silts and outwash gravel of the Ross Glacial Stage in the Southern Alps (Wellman 1951; Gage 1961). These glaciogenic strata overlie a regressive marine succession which includes Late Pliocene coquina limestone (early Waitotaran, Waipipian; Beu in Suggate et al. 1978) and lignite (early Nukumaruan, Hautawan; Mildenhall in Suggate et al. 1978). *Prima facie*, therefore, the Ross glacial deposits accumulated sometime after the c. 2.58 Ma base of the Hautawan (Fig. 7), but identifying which particular phase(s) of cooling they were associated with awaits their better dating.

Fleming (1944, 1953) described macrofaunal evidence for the northward penetration to 41°S of cold subantarctic marine waters across New Zealand in the Late Pliocene. The indicator organism, the scallop *Zygochlamys delicatula*, appears once only in the early Nukumaruan Hautawa Shellbed (2.45 Ma; MIS 97) in Wanganui Basin and in the Sentry Box Limestone at the Kuripapango Strait further north (Browne 2004), but multiple times in somewhat younger Nukumaruan sediments (2.25–1.8 Ma; MIS 87-65) in the Wairarapa Basin further east (Vella & Nicol 1970; Gammon 1997). The Hautawa Shellbed appearance has been interpreted as correlating with interglacial MIS 97 by Naish & Kamp (1995) and with MIS 95 by Cooper (2004, p. 226). Alternatively, and noting that the *delicatula* valves occur mostly towards the base of the shellbed (T. Naish pers. comm.), Wanganui Basin cycle 2 might encompass sediments deposited during both MIS 97 and preceding glacial 98, or MIS 95 and preceding glacial 96, with the glacial level being marked by the *delicatula* zone at the base of the cycle. In either case, however, major living populations of this scallop are restricted today to latitudes south of c. 45°S, so northward movement of several hundred km is indicated. The geographic distribution of the *Zygochlamys* occurrences indicates that north-travelling pulses of subantarctic water regularly reached eastern North Island from the late Pliocene onwards (cf. Nelson et al. 2000), but that only the first of these, at c. 2.45 Ma, penetrated westward through the Manawatu Straits and into Wanganui Basin, probably because tectonic uplift blocked the strait shortly thereafter (cf. Beu et al. 1997).

In the 1960s, foraminiferal studies also started to yield evidence for the occurrence of Late Pliocene cold periods. Low dextral:sinistral coiling ratios in planktic *Neogloboquadrina* *pachyderma* (Jenkins 1967, 1968) and populations of relatively large sized benthic *Nonionellina flemingi* with many chambers in their last whorl (Lewis & Jenkins 1969), both criteria that today are associated with cold waters, occur in early Waitotaran sediments at Wanganui. The possible biostratigraphic age for the three Waipipi Shellbeds overlaps with the 3.12 Ma (MIS 130) and 3.30 Ma (MIS 140) and 3.38 Ma (MIS 146) cold periods depicted in the gamma curve. More recent work, using foraminiferal censuses (modern analogue technique, MAT, of Prell 1985), has documented the occurrence of sharp Late Pliocene coolings at Site 1125 on the northern flank of the Chatham Rise (Sabaa et al. 2004). Strong surface water coolings of up to 8°C below modern occur at this site at 3.30–3.35, 2.97–3.0, and 2.79–2.82 Ma, which within the possible age model errors correspond with the cold climaxes at Site 1119 at 3.30–3.38, perhaps 3.10–3.12 and 2.78–2.80 Ma.

From the 1970s, the application of oxygen isotope analysis to foraminiferal tests delineated the occurrence of several periods of Late Pliocene cooling. Devereux et al. (1970), describing the classic Mangaopari Stream section, Wairarapa, inferred a cool interval in the base of the Greycliffs Formation, near the level of first occurrence of *Zygochlamys delicatula* (c. 2.3 Ma; MIS 87). Later work on this section by Gammon (1997) delineated also earlier periods of cooling just below the Waipipian/Mangapanian boundary at c. 3.1 Ma, and near the Waitotaran/Opoitian boundary at c. 3.6 Ma. Similar Hautawan (c. 2.3–2.4 Ma), Waipipian (c. 2.9–3.1) and late Opoitian (c. 3.5–3.6 Ma) cool periods proved to be present also in isotope records from DSDP Site 284 on Lord Howe Rise (Kennett et al. 1979; Kennett 1985).

Finally, other evidence for Late Pliocene cooling in onland New Zealand successions includes (i) the presence of a block of exotic granite inferred to have been ice rafted into the Waipipian part of the Mangapaori Mudstone, at least 200 km north of the modern limit of ice rafting (Vella 1975, p. 89), and (ii) a cool climate pollen assemblage (*Nothofagus fusca* dominant) from sediments at the Waipipian type section in Wanganui Basin (Mildenhall & Harris 1970).

Evidence for Late Pliocene warmth

Taken together, these historic data are compelling evidence for at least three New Zealand Late Pliocene coolings on a longer wavelength than the 41 k.y. Milankovitch cycle, in the Early Nukumaruan (c. 2.4 Ma), Early Waitotaran (Waipipian) (c. 3.0 Ma), and Late Opoitian (c. 3.5 Ma), respectively. However, the suggestion of Waipipian glaciation has previously provoked spirited discussion between Beu (1974) and Vella et al. (1975) regarding contradictory evidence from foraminifera and molluscs. Beu (1974), with Fleming (1953), pointed out that molluscs of warm water affinities, such as *Maoricardium*, *Crassostrea*, and *Neopanis*, are characteristic of the Waipipian shellbeds at Waipipi Beach, Wanganui. He therefore questioned the reliability of inferring cold Late Pliocene temperatures from foraminiferal evidence, concluding that "the palaeoecology of planktonic foraminifera is not well enough understood for their geographic ranges to be used as direct evidence of sea temperatures". When this conclusion was disputed by Vella et al., Beu asserted further (in Vella et al. 1975) that "there can be no doubt that the Waipipian molluscan faunas...lived in seas warmer those those around New Zealand today". Fleming & Beu's assessment of the tropical affinities of some Waipipian mollusca, and Vella et al.'s summary of the evidence for Waipipian cooling, appear to be equally correct, and it is, therefore, not surprising that this conflict of evidence has remained unresolved.

Based on this discussion, and with the benefit of the additional evidence from Site 1119, three possible explanations for this conflict can be suggested:

1. The Waipipian molluscs were living in warm bottom waters, whereas the foraminifera were introduced from pulses of subantarctic surface water (Beu in Vella et al. 1975, p. 200). Whilst this is theoretically possible, shelf waters <100 m deep, such as those from which the

Waipipi succession was mostly deposited, lie largely above the seasonal thermocline and are, therefore, generally well mixed. A stratified watermass, cold on top, also fails to account for the large sizes of the benthic foraminifer *Nonionellina* in the Waipipi succession, as described and interpreted by Lewis & Jenkins (1969).

2. The distribution of both molluscs and benthic foraminiferans are controlled strongly by seabed type (sediment facies) as well as by water depth or water temperature. For example, strong differences in faunal assemblage occur between transgressive type A shellbeds and mid-cycle type B shellbeds in the classic Castlecliff cyclothems (Abbott & Carter 1994; Abbott 1997), despite their both having been deposited in shallow water during only warm, interglacial half cycles (Beu & Edwards 1984; Beu & Kitamura 1998). Elsewhere, at deeper water locations, which preserve sediments of a full glacial/interglacial cycle, no simple pattern occurs between lithology and climate. Sand-shellbed facies (and, conversely, siltstone facies) may correspond to either glacial or interglacial periods, and, therefore, colder or warmer water, depending upon the local controls on terrigenous sediment supply (Orpin et al. 1998; Carter & Gammon 2004). This notwithstanding, the conflict between foraminiferal and molluscan evidence in the Waipipi shellbeds could perhaps be related to environmental factors other than temperature, depending upon the precise sequence stratigraphy of the samples.

No detailed sequence stratigraphic interpretation has yet been published for the Waipipi Shellbeds. However, their similarity to mid-cycle shellbeds in younger strata at Wanganui suggests their accumulation as type B shellbeds in offshore shelf waters during peak interglacial highstands. Water temperatures would then be at their warmest within a cycle, perhaps to the degree necessary to sustain the known Waipipian molluscan fauna. Should the highstand shelf waters in which the shellbeds accumulated have been deep enough to allow marine conditions to persist during ensuing glacial lowstands, then the shallow marine bioturbated sands that enclose the shellbeds might be of glacial lowstand origin. If so, these sands could have been deposited from waters that were several degrees cooler than the shellbeds, and be the source of the apparently conflicting cold water foraminiferal evidence. Bed by bed sampling and detailed sequence interpretation are needed to test this possibility.

3. The Site 1119 gamma curve, as an inferred atmospheric climate proxy, comprises an altogether new type of evidence for Late Pliocene coolings in New Zealand. Taken at face value, the gamma data indicate (i) conditions warmer than those of the Holocene during as many as 20 interglacial optima between 3.52 and 0.32 Ma (MIS 155–MIS 9); and (ii) between these warm optima, unusually cold glaciations at and around the climaxes of multi-cycles I–B, at 3.65, 3.38, 3.12, 2.72, 2.53, 1.79, 1.53, 0.52, and 0.43 Ma. (MIS Gi-2 to MIS 12) (Fig. 5F). Given the short period over which some of these warm optima and cold climaxes alternate in the Late Pliocene part of the record, the conflicting molluscan and foraminiferal evidence from onland sections might be explained by samples having been drawn from slightly different horizons and representing closely adjacent warm and cold periods. As for interpretation 2, detailed sampling and study of relevant onland Waipipian sequences are required to test and elaborate this idea.

Implications for faunal history

Assuming a linear relationship between gamma ray value and temperature (Carter & Gammon 2004), the Site 1119 warm optima at 3.52, 3.16, 2.9, and 2.4 Ma had mean temperatures 2–4°C warmer than modern values. Transferred to Wanganui, such enhanced temperatures would be adequate to explain the presence of warm water molluscs in interglacial shellbeds up until 2.4 Ma (MIS 97; the Hautawa Shellbed, which contains the last known occurrence of several warm water genera). Second, the last warm optimum before the overall colder Late

Pleistocene occurred at 0.95 (MIS 25). This is close to the 1.1 Ma date estimated by Beu (1990, p. 286) for the otherwise enigmatic disappearance from the fossil record of 13% of New Zealand molluscan genera around the end of Nukumaruan time, and also corresponds with a small, but notable, influx of warm water immigrants including *Maoricrypta costata*, *Iredalula striata*, and *Bembicium auratum* (Beu 2004). Third, the Site 1119 gamma curve, and a foraminiferal census analysis of SST at Sites 594 and 1119 (Wilson et al. 2005), suggests that warmer than usual interglacial temperatures occurred most recently at 0.42 Ma (MIS 11), 0.32 Ma (MIS 9), and 0.125 Ma (MIS 5e). This evidence is consistent with the presence of *Anadara trapezia* and a significant number of other warm water molluscan immigrants to New Zealand waters at these times (Murray-Wallace et al. 2000; Beu 2004).

That several known faunal variations occur in a manner consistent with past temperatures predicted from the Site 1119 gamma curve suggests that shallow marine molluscan assemblages in New Zealand respond sensitively to moderate mean temperature changes of 2–4°C. Another important corollary of the discerned climatic pattern is that throughout the rapidly fluctuating G/I fluctuations of the Late Pliocene and Pleistocene, stenothermal species of tropical and subantarctic affinity will only have been present during periodic warm optima or cold climaxes, respectively, and will therefore possess discontinuous stratigraphic ranges (cf. Beu 2004, fig. 5). Given also that the New Zealand landmass ranges across 14° of latitude, the first and last occurrences of species ranges will also have differed from place to place across latitude, as benthic assemblages reacted to environmental conditions controlled both by local geography, itself often changing because of tectonic effects, and an oscillating global climate.

Summary

The ages estimated for onland Late Pliocene coolings older than 2.5 Ma have in the past been based largely on biostratigraphy. Therefore, they may have an error of up to several hundred thousand years, and not yet be correlated accurately between different datasets or regions. Nonetheless, the Site 1119 gamma data adds to what is already very strong evidence for older Late Pliocene coolings in the New Zealand region. *Prima facie*, the matches between earlier evidence and the multi-cycle coolings apparent at Site 1119 are as follows:

	Ma	MIS	Site 1119 multi-cycle
Fleming (1944)	2.4	97/96	E/F boundary
Jenkins (1967)	2.8–2.9	114–120	F = cold climax
Lewis & Jenkins (1969)	2.8–2.9	114–120	F = cold climax
Devereux et al. (1970)	2.3	88	E = cold climax
Kennett et al. (1971, 1979)	3.6	156	I = cold climax
Gammon (1997)	3.1	130	G = cold climax
	3.6	Gi-2	I = cold climax

Causes of the Late Pliocene natural gamma ray maxima

Recalling that the 1119 gamma curve reflects runoff from eastern South Island rivers, three interpretations are possible for the Late Pliocene "cooling" excursions seen in the curve. First, these features may represent real climatic coolings of up to 12°C, accompanied by mountain glacier formation. The presence of strong signals in the New Zealand gamma ray data compared with their weak reflection in global ice volume isotope curves would then imply that these events were amplified in the mid-latitudes, perhaps because of changes in precipitation driven by latitudinal shifts in the northern edge of the zonal westerly wind system (Fitzharris et al. 1992). Alternatively, and second, Pliocene glacier development in the Southern Alps, and mud

(= enhanced gamma ray) yields, may have been driven not so much by climatic cooling as by the occurrence of phases of orogenic uplift of mountains to heights at which glaciers formed, an idea that goes back to Haast (1864). Stratigraphic evidence for Pliocene mountain building associated with plate collision has been documented at three locations along the Alpine Fault (c. 5.0 Ma, Cutten 1979; c. 5.0–4.0 Ma, Sutherland 1996; c. 3.5–3.0 Ma, Mortimer et al. 2001). With due allowance for varying transport paths and periods towards the east, these dates are consistent with the sharp rise in chlorite-illite-dominated clay assemblages that occurs east of South Island between 3.5 and 2.4 Ma at Site 594 (Dersch & Stein 1991) and 3.5–2.0 Ma at Site 1119 (Winkler & Dullo 2002) (see also Carter et al. 2004c, fig. F18). Third, and alternatively again, the clay mineral assemblage changes and the Late Pliocene gamma highs at Site 1119 might have been driven by tectonics in their own right rather than through a climate (ice) link, as Pliocene uplift exposed formerly buried, higher grade metamorphic rocks to release abundant biotite, sericite and orthoclase into eastern South Island watersheds. According to such an interpretation, parts of the Late Pliocene gamma record (especially any longer term trends), could then be a proxy signal for hinterland tectonism rather than climate *per se*.

Evidence for Late Pliocene coolings elsewhere

The mid–late Pliocene cooling episodes represented at Site 1119, and elsewhere in the New Zealand region, are not just a regional phenomenon. For example, Anderson (1997) has recorded an extended cool period between 3.35 and 3.05 Ma at ODP Site 806 in the western equatorial Atlantic Ocean; Site 606 in the North Atlantic has isotope enrichment excursions at 3.2, 2.7, 2.6, and 2.4 Ma (Keigwin 1986); Heusser & Morley (1996) document mid-Pliocene coolings from the vicinity of Japan, western Pacific Ocean, based on combined pollen and microfaunal study; St John (2004) has demonstrated the presence of major pulses of ice rafted detritus off Greenland and Antarctica at 3.5 Ma (Greenland only) and 2.9–2.7 Ma; and Marlow et al. (2000), using alkenone calculated SSTs, showed the presence of marked Pliocene coolings at 3.7, 3.2, 2.8, and 2.5 Ma at ODP Site 1084 off Namibia. These pulsatory coolings, and iceberg flotilla events, took place against a background of global mid-Pliocene warmth (Dowsett et al. 1996).

Because the Site 1119 coolings are inferred from a proxy record that is related to mid-latitude glacial intensity, they are relevant to the long standing dispute over deglaciation of Antarctica in the middle Pliocene. Protagonists in this debate assert that Pliocene diatoms found in tills of the Sirius Formation in the Transantarctic Mountains are either *in situ*, representing periods

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Fig. 7 *Inserted between pages 26 and 27:* Climatic reference curve for New Zealand for the last 3.9 Ma, in 1.0 m.y. long segments. Depicted from the top down for each segment is: the Geomagnetic Polarity Scale (GPS) (after Lourens et al. 1996; Shackleton et al. 1995; Horng et al. 2002; Gradstein et al. 2004); the planktic/benthic oxygen isotope stage record and MIS assignations from ODP Site 849 (Shackleton et al. 1995); numbered basin cyclothems from Wanganui Basin (after Saul et al. 1999); terminology for New Zealand stages/oppelzones, glacial stages and Wanganui coastal terraces (after Carter & Naish 1998; Suggate 1965, 1990; Pillans 1990); varied correlation indicators, keyed into the timescale by upward pointing arrows (for published sources, see right-hand column in Table 1); natural gamma ray composite time series from ODP Site 1119 (after Carter & Gammon 2004); + signs along top margin indicate timescale tie-points to Vostok and ODP Site 758; black upward-pointing pyramids along bottom margin indicate downhole depths in increments of 50 rmcd); sand log for Site 1119 (after Carter & Gammon 2004); core retrieval (black: core retrieved; white: core lost); and carbonate percentage timeseries for Site 1119, calculated by transformation of core reflectance measurements (after Millwood et al. 2002). *Addendum added in proof:* The paper by Pillans et al. 2005 this issue) introduces a modified numbering scheme for the Wanganui Basin cyclothems. Above the Pakihikura tephra (MIS 51), this renumbering results in shifts of up to two cycles in the correlation of the Saul et al. (1999) units with the standard MIS (T. Naish pers comm.)

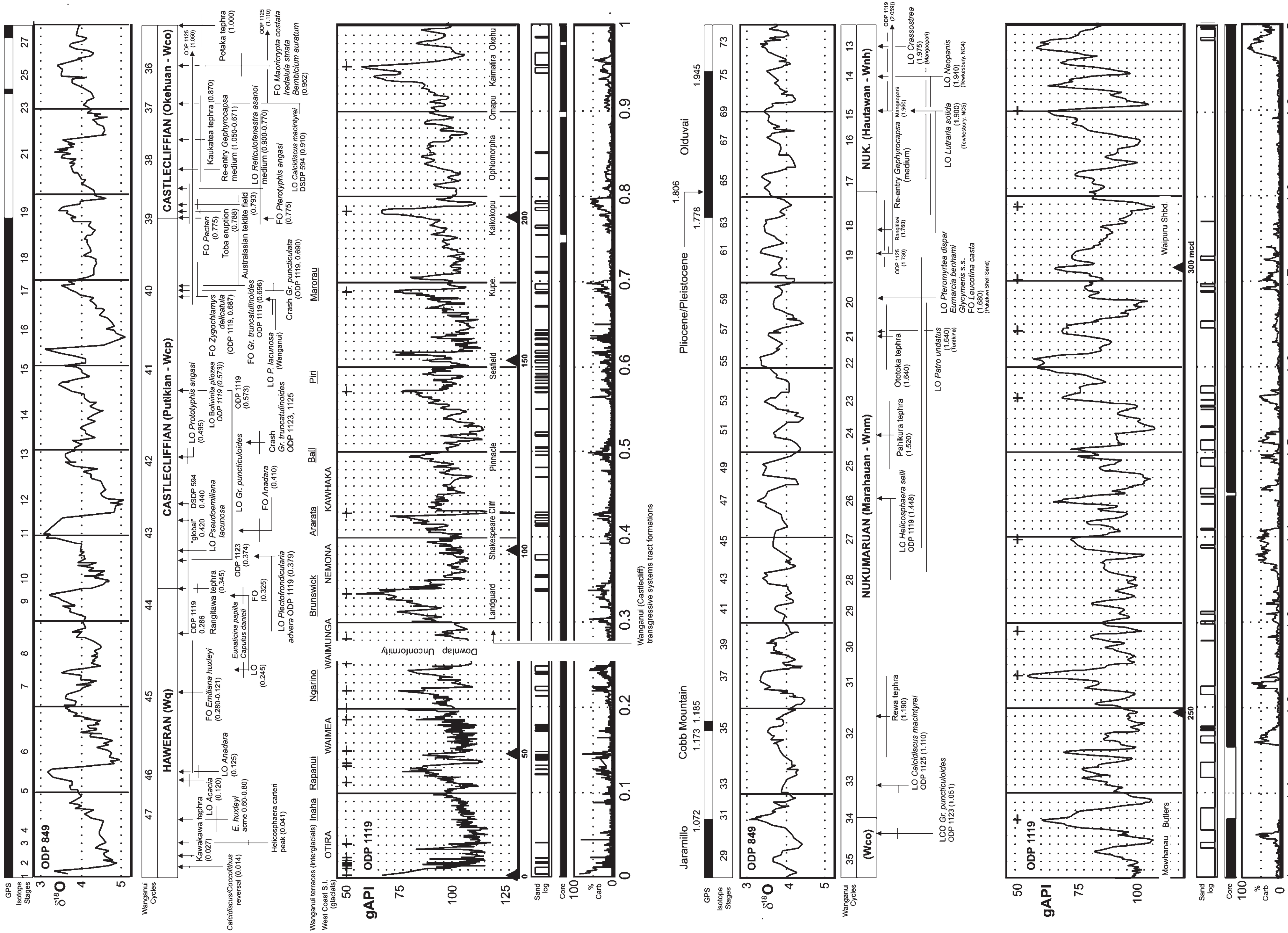

Fig. 7 *Caption on page 26*

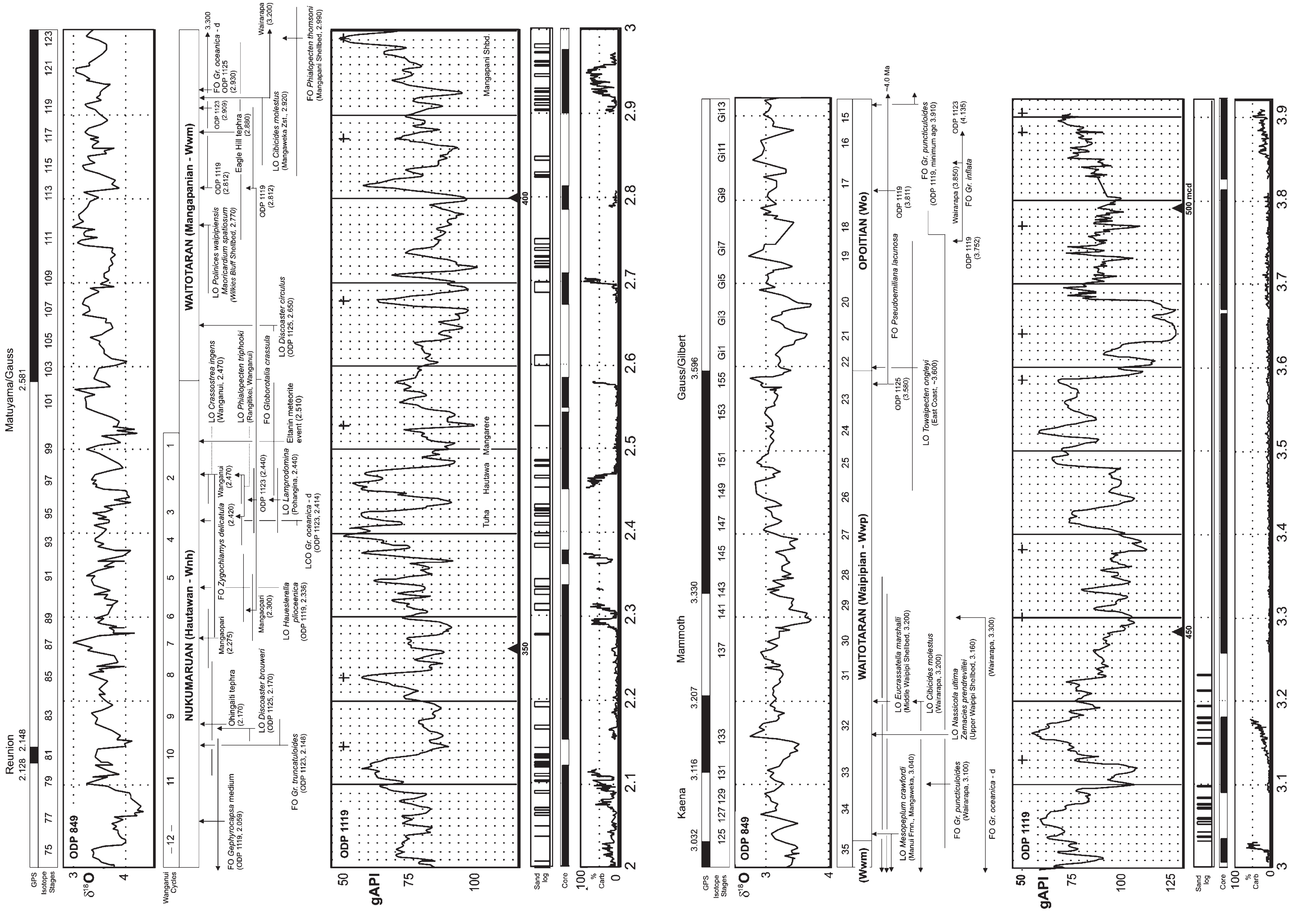

Fig. 7 *Caption on page 26*

of warming and deglaciation (Webb et al. 1984), or instead are wind-blown contaminants (Stroeven et al. 1996). The occurrence of strong cooling-warming cycles of both sudden (3.68–3.59 and 3.16–3.06 Ma) and more gradual (3.52–3.16 and 2.99–2.40 Ma) nature within multi-cycles I–F at Site 1119 (Fig. 5F) is consistent with alternating mid-latitude glaciation and deglaciation during the middle Pliocene. Given the extremely close match that has been demonstrated for atmospheric climate cyclicity between New Zealand and Antarctica during the Late Pleistocene (Carter et al. 2004c; Carter & Gammon 2004), it is, therefore, highly probable that climatic cycling similar to that of the New Zealand mid-Pliocene occurred in Antarctica also. Thus, the data from Site 1119 favour the "dynamicist" interpretation of Pliocene climate in Antarctica.

SITE 1119 AS A STRATIGRAPHIC TEMPLATE

A glance at any map of New Zealand (e.g., Fig. 1) reminds us, first, that the country straddles 14° of the middle latitudes; and, second, that the emergent islands, lying athwart the cool subtropical-warm subantarctic wind belts, oceanic fronts, and ocean waters, are situated in a dynamic environmental setting. The wide latitudinal range has a marked effect on the zoogeographic distribution of organisms used in correlation, such as fossil molluscs and foraminifers. Given this, and given also the climatic changeability that has accompanied the last few million years of Earth history, not to mention the overprint of plate boundary tectonics, the correlation of Pliocene-Pleistocene strata in New Zealand is unlikely to be straightforward. Shifting fronts and watermasses, shifting atmospheric wind belts, shifting basin terranes and shifting temperatures all militate against the simple assumption of synchrony of appearance or disappearance that underpins traditional biostratigraphic correlation. As discussed in more detail by Carter (in press), and summarised in Table 1 and Fig. 7, very few if any fossil datums can be used to date Pliocene-Pleistocene sediments accurately countrywide. Therefore, other methods, such as magnetostratigraphy, isotope stratigraphy, tephrostratigraphy, and cyclostratigraphy, have recently come to assume central importance for correlation of New Zealand's younger sedimentary sequences (cf. Table 1).

Against such a background, no one stratigraphic section can represent every facet of New Zealand's geological history throughout the late Cenozoic ice ages. That acknowledged, there remains the need for a standard record against which data of many different types and from many different sources can be assessed. The natural gamma ray record from Site 1119 has a resolution of 1–2 k.y. back to 3.9 Ma (Carter & Gammon 2004), and is currently the most suitable record for use as a regional climatic template (Fig. 7). Because the site is situated in the marine environment, it contains a record of changing oceanographic (Carter et al. 2004a,b; Wilson et al. 2004), marine biostratigraphic (Hayward in Carter et al. 1999), and potentially terrestrial vegetation events (cf. Heusser & de Geer 1994; Mildenhall 2003). At the same time, because the gamma ray signal acts as a close proxy for climatic cycling in onland South Island, and more widely to at least Antarctica (Carter & Gammon 2004), the Site 1119 record reflects also the changing ice volumes and intensity of glaciation in the Southern Alps, which is an atmospheric signal. Finally, located as it is c. 100 km offshore, and on the upper continental slope, Site 1119 serves as an important bridge between more landward records of marine shelf and terrestrial nature and more seaward cores which contain a deep water record of AAIW (DSDP 594; Nelson et al. 1993), CDW (ODP Site 1123; Hall et al. 2001) and global (Nelson et al. 1985, 1993) climatic history.

Figure 7 comprises the composite natural gamma record for Site 1119 plotted at high resolution and compared with a parallel oceanic oxygen isotope record from ODP Site 849 in the eastern equatorial Pacific Ocean (Shackleton et al. 1995). Significant events and features, especially those of use for regional correlation, are displayed between the oxygen isotope and

Table 1 Table of correlation criteria for the last 3.9 Ma in New Zealand and the surrounding south-west Pacific Ocean region. FO, first occurrence; LO, last occurrence; LCO, last common occurrence; n, nannoplankton; p, plant; f, foraminifera; m, mollusca; M, magnetic chron; T, tephra; a, astro-event. Cycle numbering after Saul et al. (1999; with the addition of a prefix W to signal Wanganui Basin), Journeaux et al. (1996), McIntyre & Kamp (1998), and Abbott et al. (2005 this issue). FO and LO levels from different stratigraphic sections are numbered arbitrarily, e.g., FO-1 to FO-n, in order of decreasing age. 3.91 Ma is listed as a minimum FO age for *Gr. punticuloides* and *Gr. inflata* at Site 1119 because they first appear at the very bottom of the core, leaving open the possibility that their true FO is older. The error estimates in this table are mostly qualitative, and many are based upon an assessment of the likelihood of the event being miscorrelated to the MIS scale by one (40 k.y.), two (80 k.y.) or more cycles. The ages given for tephra are those estimated from the age of the MIS cycle within which they appear to occur; the best available numeric date, and error, for each is listed in the adjacent (error) column.

Criterion		Code	mbsf	(r)mcd/cycle	Location	Age (m.y.)		Error (k.y.)	MIS	References
Calcidiscus/Coccolithus	reversal	n			Q858/DSDP 594/ODP 1123	0.014	±	1	2/1	Fenner & Di Stefano (2004)
Calcidiscus/Coccolithus	reversal	n	1.71	1.75	ODP 1119 (B1H-2-21)	0.014	±	1	2/1	Fenner & Di Stefano (pers. comm.)
Kawakawa tephra	eruption	T		W Cycle 47	Taupo Volcanic Zone	0.027	±	2	3	Alloway et al. (2004)
Helicosphaera carteri	peak	n	17.96	19.20	ODP 1119 (B3H-3-76)	0.041	±	3	3	Fenner & Di Stefano (pers. comm.)
Emiliana huxleyi-2	acme-2	n	5.58		DSDP 594 (1H-4-8)	0.060	±	7.5	4/3	Wei & Chen (pers. comm.)
Gephyrocapsa medium	crash	n			Q858/DSDP 594/ODP 1123	0.060	±	5	4/3	Fenner & Di Stefano (2004)
Emiliana huxleyi-1	acme-1	n	1.56	1.56	ODP 1125 (A1H-2-6)	0.080	±	7.5	5	Wei & Chen (pers. comm.)
Acacia	LO	p		W Cycle 46	Rapanui Lignite, Wanganui	0.120	±	10	5	Bussell & Mildenhall (1990)
Emiliana huxleyi-4	FO-4	n	43.58	44.89	ODP 1119 (B5H-cc)	0.121		+30	5	Wei in Carter et al. (1999)
Anadara trapezia	LO	m			Many North I. last interglacial terraces	0.125	±	10	5	Beu (2004)
Reticulofenestra pseudoumbilicus	LO	n	72.09	77.39	ODP 1119 (B8H-cc)	0.195		–15	7	Wei in Carter et al. (1999)
Emiliana huxleyi-3	FO-3	n	22.4		DSDP 594 (3H-5-9)	0.230		+15	7	Wei & Chen (pers. comm.)
Emiliana huxleyi-2	FO-2	n		W Cycle 45	Waipuna Bay Fmn, Wanganui	0.245	±	25	7	Naish et al. (1998)
Eunaticina, Capulus spp.	LO	m			Te Piki fossil beds, East Cape	0.245	±	25	7	Beu 2004
Emiliana huxleyi-1	FO-1	n	4.56	5.35	ODP 1125 (A2H-1-26)	0.280		+7.5	8	Wei & Chen (pers. comm.)
Pseudoemiliana lacunosa-5	LO-5	n	81.5	87.51	ODP 1119 (B9H-cc)	0.286		–7.5	9	Wei in Carter et al. (1999)
Eunaticina, *Pupa*, *Capulus* spp.	FO	m		W cycle 44	Landguard Sand, Wanganui	0.335	±	10	9	Beu (2004)
Rangitawa tephra	eruption	T		W Cycle 43/44	(= base Haweran)	0.340		0.345 ± 12	10/9	Pillans et al. (1996)
Globorotalia puncticulata	LO	f		c. 95.00	ODP 1119	c.0.352	±	10	10	Scott et al. (in prep.)
Globorotalia puncticuloides-3	LO-3	f	13.90	15.62	ODP 1123 (B3H-1-100)	0.374	±	5	10	Scott (pers. comm.)
Plectofrondicularia advera	LO	f	90.93	99.13	ODP 1119 (B10H-cc)	0.379		–100	10	Hayward in Carter et al. (1999)
Anadara trapezia	FO	m		W Cycle 43	Rangitawa fossil beds, Rangitikei	0.410	±	15	11	Beu (2004)
Globorotalia puncticulata-r	re-entry	f		c. 105	ODP 1119	c.0.419	±	10	11	Scott et al. (in prep.)
Pseudoemiliania lacunosa-4	LO-4	n			"global"	0.420			12/11	Sato & Kameo (1996)
Pseudoemiliania lacunosa-3	LO-3	n	50.67		DSDP 594 (6H)	0.440		–15	12	Wei & Chen (pers. comm.)
Pseudoemiliania lacunosa-2	LO-2	n	7.31	8.88	ODP 1125 (A2-3-1)	0.460		–7.5	12	Wei & Chen (pers. comm.)
Prototyphis angasi	LO	m			Tainui Shellbed, Wanganui	0.495	±	10	13	Beu (2004)
Globorotalia truncatulinoides-d	crash	f		9.45	ODP 1125	0.502	±	15	13	Hayward (pers. comm.)

Globorotalia truncatulinoides-d	crash	f	21.40	23.12	ODP 1123 (B3H-6-100)	0.520	±	15	14/13	Scott et al. (in prep.)
Bolivinita pliozea	LO	f	120.84	133.00	ODP 1119 (B14H-2-64)	0.573		–100	14/13	Wilson et al. (2004)
Globorotalia puncticuloides-2	LO-2	f	120.84	133.00	ODP 1119 (B14H-2-64)	0.573		–100	14/13	Scott (pers. comm.)
Gephyrocapsa medium-4	re-entry-4	n	160.96	179.06	ODP 1119 (C17H-cc)	0.671		+100	16	Wei in Carter et al. (1999)
Pseudoemiliania lacunosa-1	LO-1	n		W Cycle 40	Upper Kai-iwi Siltstone, Wanganui	0.680		–20	17/16	Beu & Edwards 1984
Zygochlamys delicatula-3	FO-3	m	168.00	186.75	ODP 1119 (C18X-6)	0.687		+40	17	Carter et al. (1999)
Globorotalia puncticulata	crash	f	168.53	187.28	ODP 1119 (C18X-7-28)	0.690		+12	18	Scott et al. (in prep.)
Globorotalia truncatulinoides-2	FO-2	f	169.38	188.13	ODP 1119 (C18X-cc)	0.696		+100	18/17	Hayward in Carter et al. (1999)
Reticulofenestra asanoi-3	LO-3	n		W Cycle 39	Upper Westmere Zst., Wanganui	0.770		+40	19	Naish et al. (1998)
Pecten s.s.	FO	m		W Cycle 39	Kaikokopu Shellbed, Wanganui	0.775		+40	19	Fleming (1953)
Prototyphis angasi	FO	m		W Cycle 39	Kaikokopu Shellbed, Wanganui	0.775		+40	19	Beu (2004)
Early/Late Pleistocene boundary	reversal	M			global	0.781	±	1.7	19	Pillans (2003)
Matuyama/Brunhes boundary	reversal	M		W Cycle 39	Kaikokopu Shellbed, Wanganui	0.781	±	1.7	19	Turner & Kamp 1990; Pillans et al. (1994)
Base Brunhes Chron	reversal	M			(= base Putikian)	0.781	±	1.7	19	Lourens et al. in Gradstein et al. (2004)
Toba super-eruption	eruption	T			ODP 758, South China Sea	0.788	±	2.2	20/19	Lee et al. (2004)
Australasian tektite field	impact	a			MD-972142 /3, South China Sea	0.793	±	5	20	Lee & Wei (2000)
Reticulofenestra asanoi-2	LO-2	n			"global"	0.850			21	Sato & Kameo (1996)
Kaukatea tephra	eruption	T		W Cycle 37	Taupo Volcanic Zone	0.870		0.87 ± 50	23	Shane et al. (1996)
Reticulofenestra asanoi-1	LO-1	n	18.31	18.20	ODP 1125 (A3H-5-1)	0.900		–7.5	23	Wei & Chen (pers. comm.)
Calcidiscus macintyrei-2	LO-2	n	87.75		DSDP 594 (10H)	0.910		–15	23	Wei & Chen (pers. comm.)
Maoricrypta, *Iredalula*, *Bembicium* spp.	FO	m			Kaimatira Pumice Sand, Wanganui	0.952	±	15	25	Beu (2004)
Jaramillo subchron	top	M		W Cycle 36/37	Below Kaukatea tephra	0.988	±	5	27	Pillans et al. (1994); Naish et al. (1996)
Jaramillo subchron	top	M			global	0.988	±	5	27	Lourens et al. in Gradstein et al. (2004)
Potaka tephra	eruption	T		W Cycle 36	Taupo Volcanic Zone	1.000		1.00 ± 30	27	Alloway et al. (in press)
Gephyrocapsa (medium)-3	re-entry-3	n	98.85		DSDP 594 (11H-4-55)	1.030		+15	28	Wei & Chen (pers. comm.)
Gephyrocapsa (medium)-1	re-entry-1	n	22.54	22.45	ODP 1125 (A3H-6-124)	1.050		+7.5	30	Wei & Chen (pers. comm.)
Globorotalia puncticuloides-1	LCO	f	37.40	43.02	ODP 1123 (B5H-4-100)	1.051	±	5	30	Scott (pers. comm.)
Jaramillo subchron	reversal	M		base W Cycle 35	Okehu Zst., inland Wanganui	1.072	±	5	31	Pillans et al. (1994); Naish et al. (1996)
Base Jaramillo subchron	reversal	M			(= base Castlecliffian, Okehuan)	1.072	±	5	31	Lourens et al. in Gradstein et al. (2004)
Calcidiscus macintyrei-2	LO-2	n	23.90	25.39	ODP 1125	1.110		–7.5	32	Wei & Chen (pers. comm.)
Cobb Mountain subchron	reversal	M		W Cycle 32	Above Rewa tephra, Rangitikei Valley	1.173	±	12	35	Pillans et al. (1994); Naish et al. (1998)
Cobb Mountain subchron	top	M			global	1.173	±	5	35	Lourens et al. in Gradstein et al. (2004)
Cobb Mountain subchron	base	M			global	1.185	±	5	35	Lourens et al. in Gradstein et al. (2004)

Table 1 (*continued*)

Criterion		Code	mbsf	(r)mcd/cycle	Location	Age (m.y.)		Error (k.y.)	MIS	References
Rewa tephra	eruption	T		W Cycle 32	Taupo Volcanic Zone	1.190		1.08 ± 60	35	Alloway et al. (in press)
Helicosphaera selli	LO	n	254.63	273.38	ODP 1119 (C27X-cc)	1.448		–100	47	Wei in Carter et al. (1999)
Pahikura tephra	eruption	T		W Cycle 24	Taupo Volcanic Zone	1.520		1.58 ± 50	51	Shane et al. (1996)
Calcidiscus macintyrei -1	LO-1	n			“global”	1.600			55	Raffi et al. (1993)
Patro undatus	LO	m		W Cycle 21	Turakina section	1.640		–300	57	Naish et al. (1998)
Ototoka tephra	eruption	T		W Cycle 21	Taupo Volcanic Zone	1.640		1.71 ± 0.19	57	Alloway et al. in Beu (2004)
Pteromyrtea, *Eumarcia*, *Glycymeris* spp.	LO	m		W Cycle 20	Pukekiwi Shell Sand (NC-10)	1.680		–300	59	Beu (2004)
Leucotina casta	FO	m		W Cycle 20	Pukekiwi Shell Sand (NC-10)	1.680		+50	59	Beu (2004)
Gephyrocapsa (medium)-4	FO-4	n	37.31	42.36	ODP 1125 (A5H-3-141)	1.730		+7.5	61	Wei & Chen (pers. comm.)
Gephyrocapsa sinuosa-3	FO-3	n		W Cycle 18	Rangitikei Valley, Wanganui	1.760	±	40	63	Naish et al. (1998)
Olduvai subchron	reversal	M		base W Cycle 18	Waipuru Shellbed, Rangitikei Valley	1.778	±	5	63	Seward et al. (1986); Naish et al. (1996)
Olduvai subchron	top	M			Global	1.778	±	5	63	Lourens et al. in Gradstein et al. (2004)
Pliocene/Pleistocene boundary				W Cycle 17	Vinegar Hill Fmn (Vhzm-2)	1.806	±	10	65	Naish et al. (1996, 1997)
Pliocene/Pleistocene boundary					(= base Marahauan)	1.806	±	10	65	Lourens et al. (1997)
Lutraria solida	LO	m		W Cycle 15	Tewkesbury Fmn (NC-5)	1.900		–150	69	Abbott et al. (2005 this issue)
Gephyrocapsa sinuosa-2	FO-2	n		Cycle P4	Pukenui Limestone, Mangaopari	1.900		+80	69	Gammon (1997)
Neopanis neozelanica	LO	m		W Cycle 14	Tewkesbury Fmn (NC-4)	1.940		–150	71	Beu (2004); Abbott et al. (2005 this issue)
Olduvai Subchron	reversal	M		base Cycle 14	(inferred only)	1.945		–8	71	Seward et al. (1986); Naish et al. (1996)
Olduvai Subchron	base	M			Global	1.945		–8	71	Lourens et al. in Gradstein et al. (2004)
Crassostrea ingens-2	LO-2	m		Mangaop. Cycle P1	Pukenui Limestone, Mangaopari	1.975		–80	73	Gammon (1997); Beu & Maxwell (1990)
Gephyrocapsa (medium)-1	FO-1	n	312.83	331.58	ODP 1119 (33X-cc)	2.059		–100	78	Wei in Carter et al. (1999)
Reunion Subchron	top	M		W Cycle 10	Not yet identified in onland New Zealand	2.128	±	5	81	Lourens et al. in Gradstein et al. (2004)
Reunion Subchron	base	M		W Cycle 10	Not yet identified in onland New Zealand	2.148	±	5	81	Lourens et al. in Gradstein et al. (2004)
Globorotalia truncatulinoides-1	FO-1	f	73.90	81.22	ODP 1123 (B9H-3-100)	2.148		–40	82	Scott (pers. comm.)
Discoaster brouweri	LO	n	44.52	52.01	ODP 1125 (A6H-2-72)	2.170		+7.5	82	Wei & Chen (pers. comm.)
Ohingaiti tephra	eruption	T		W Cycle 9	Taupo Volcanic Zone	2.170	±	80	82	Naish et al. (1998)
Zygochlamys delicatula-2	FO-2	m		Mangaop. Cycle M9	Mangaopari Mudstone	2.275	±	80	87	Gammon (1997)

Globorotalia crassula-5	FO-5	f		Mangaop. Cycle M5	Mangaopari Mudstone	2.300	±	40	89	Gammon (1997)
Haueslerella plioceneenica	LO	f	338.78	357.53	ODP 1119 (C36X-cc)	2.336	±	40	90	Hayward in Carter et al. (1999)
Globorotalia crassula-4	FO-4	f			Waiararapa (regional)	2.400	±	100	93	Shane et al. (1995)
Globorotalia oceanica-d-2	LCO-2	f	80.60	88.28	ODP 1123 (B10H-1-120)	2.414		–7	95	Scott (pers. comm.)
Phialopecten tripkooki	LO	m		W Cycle 3	Rangitikei Valley, Wanganui	2.420		–40	95	Naish et al. (1998)
Globorotalia oceanica-d-1	LO-1	f		c. 51.00	ODP 1125	2.440	±	40	96	Sabaa et al. (2004)
Lamprodomina neozelanica	LO	m		W Cycle 2/3	Basal Komako Fmn	2.440	±	80	96	Carter (1972)
Globorotalia crassula-3	FCO-3	f	82.38	89.06	ODP 1123 (B10H-2-48)	2.444		+8	96	Scott (pers. comm.)
Globorotalia crassula-2	FO-2	f			Rangitikei Valley, Wanganui	2.450	±	80	96	Naish et al. (1998)
Phialopecten tripkooki	LO	m		W Cycle 2	Hautawa Shellbed, Wanganui	2.470		–40	97 or 95	Fleming (1953); Beu (1995)
Zygochlamys delicatula-1	FO-1	m		W Cycle 2	Hautawa Shellbed, Wanganui	2.470		+40	97 or 95	Naish et al. (1998); Cooper (2004)
Crassostrea ingens-1	LO-1	m		W Cycle 2	Hautawa Shellbed, Wanganui	2.470		–40	97 or 95	Naish et al. (1998); Cooper (2004)
Eltanin meteorite event	impact	a		W Cycle 1	Mangarere basal breccia-cglt.	2.500		–40	100	Naish & Kamp (1995)
Eltanin meteorite event	impact	a			south-east Pacific	2.510	±	70	100	Frederichs et al. (2002)
Gauss/Matuyama boundary	reversal	M		nr base Cycle M11	70 m below top of Mangaweka Mudstone	2.581		–23	103	Journeaux et al. (1996); Naish et al. (1996)
Base Matuyama Chron	reversal	M			(= base Nukumaruan, Hautawan)	2.581		–23	103	Lourens et al. in Gradstein et al. (2004)
Discoaster surculus	LO	n	54.82	64.80	ODP 1125 (B6H-6-102)	2.650		–7.5	106 (G-2)	Wei & Chen (pers. comm.)
Polinices waipipiensis	LO	m		Cycle M3	Wilkies Bluff Shellbed, Wanganui	2.770	±	80	111 (G-7)	Fleming (1953); McIntyre & Kamp (1998)
Maoricardium spatiosum	LO	m		Cycle M3	Wilkies Bluff Shellbed, Wanganui	2.770	±	80	111 (G-7)	Fleming (1953); McIntyre & Kamp (1998)
Globorotalia crassula-1	FCO-1	f	382.24	400.99	ODP 1119 (C41X-cc)	2.812		+100	113 (G-11)	Hayward in Carter et al. (1999)
Globorotalia oceanica-d-5	FO/LO-5	f	382.24	400.99	ODP 1119 (C41X-cc)	2.812		+100	113 (G-11)	Hayward in Carter et al. (1999)
Eagle Hill tephra	eruption	T		Cycle J7	Taupo Volcanic Zone	2.880	±	80	117 (G-13)	Naish et al. (1998)
Cibicides molestus-2	LO-2	f		Cycle J6	Cycle 6, Mangaweka Zst	2.920		–80	119 (G-15)	Journeaux et al. (1996)
Globorotalia oceanica-d-4	FO-4	f	97.35	105.31	ODP 1123 (B11H-6-95)	2.909		+7.5	119 (G-15)	Scott (pers. comm.)
Globorotalia oceanica-d-3	FO-3	f		c. 72.00	ODP 1125 (B7)	2.930	±	40	120 (G-16)	Sabaa et al. (2004)
Globorotalia oceanica-d-2	FO-2	f		Cycle J5	Cycle 5, Mangaweka Zst.	2.950		+80	121 (G-17)	Journeaux et al. (1996)
Phialopecten thompsoni	FO	m		base Cycle J5	Mangapani Shellbed, Wanganui	2.990	±	80	125 (G-21)	Journeaux et al. (1996); Beu (2000)
Kaena Subchron	reversal	M		upper Cycle J4	upper Manui Fmn, Mangaweka	3.032		–8	125 (G-21)	Seward et al. (1986); Journeaux et al. (1996)
Top Kaena Subchron	reversal	M			(= base Mangapanian)	3.032		–8	125 (G-21)	Lourens et al. in Gradstein et al. (2004)
Mesopeplum crawfordi	LO	m		Cycle J4	Manui Fmn, Mangaweka	3.040		–200	125 (G-21)	Journeaux et al. (1996)
Globorotalia puncticuloides-2	FO-2	f			Wairarapa (regional)	3.100	±	100	130 (K-2)	Shane et al. (1995)

Table 1 (*continued*)

Criterion		Code	mbsf	(r)mcd/cycle	Location	Age (m.y.)		Error (k.y.)	MIS	References
Kaena Subchron	base	M		Cycle J3/J4	Kawhatau/Manui contact, Mangaweka	3.116	±	6	131 (KM-1)	Seward et al. (1986); Journeaux et al. (1996)
Kaena Subchron	base	M			Global	3.116	±	6	131 (KM-1)	Lourens et al. in Gradstein et al. (2004)
Nassicola ultima	LO	m			Upper Waipipi Shellbed, Wanganui	3.160	±	200	133 (KM-3)	Fleming (1953)
Zemacies prendrevillei	LO	m			Upper Waipipi Shellbed, Wanganui	3.160	±	200	133 (KM-3)	Beu & Maxwell (1990)
Eucrassatella marshalli	LO	m			Middle Waipipi Shellbed, Wanganui	3.200	±	200	135 (KM-5)	Fleming (1953)
Cibicides molestus-1	LO-1	f			Wairarapa (regional)	3.200		100	135 (KM-5)	Shane et al. (1995)
Mammoth Subchron	reversal	M		upper Cycle J3	Kawhatau Fmn, Mangaweka	3.207		–13	135 (KM-5)	Seward et al. (1986); Journeaux et al. (1996)
Mammoth Subchron	top	M			Global	3.207		–13	135 (KM-5)	Lourens et al. in Gradstein et al. (2004)
Globorotalia oceanica-d-1	FO-1	f			Wairarapa (regional)	3.300		+100	140 (M-2)	Shane et al. (1995)
Mammoth Subchron	reversal	M		nr base Cycle J2	Tarere Fmn, Mangaweka	3.330	±	0	142 (MG-2)	Seward et al. (1986); Journeaux et al. (1996)
Mammoth Subchron	base	M			Global	3.330	±	0	142 (MG-2)	Lourens et al. in Gradstein et al. (2004)
Pseudoemiliana lacunosa-3	FO-3	n	89.42	102.74	ODP 1125 (B10H-4-25)	3.580		+15	155	Wei & Chen (pers. comm.)
Gilbert/Gauss boundary	reversal	M			50 m below top of Taihape Mudstone	3.596	±	16	156	Seward et al. (1986)
Base Gauss Chron	reversal	M			(= base Waitotaran, Waipipian)	3.596	±	16	156	Lourens et al. in Gradstein et al. (2004)
Towaipecten ongleyi	LO	m			East Coast North I. (regional)	3.600	±	150	156	Beu (1995); Naish et al. (1998)
Reticulofenestra pseudoumbilicus-3	LO-3	n			Wanganui (regional)	3.620	±	100	Gi-1	Naish et al. (1999)
Reticulofenestra pseudoumbilicus-2	LO-2	n			Wairarapa (regional)	3.750	±	100	Gi-7	Shane et al. (1995)
Globorotalia inflata (3 chambers)*-3*	FO-3	f	475.65	495.40	ODP 1119 (C50X-cc)	3.752	±	80	Gi-7	Scott (pers. comm.)
Pseudoemiliana lacunosa-2	FO-2	n	483.56	502.31	ODP 1119 (51X-cc)	3.811		+100	Gi-9	Wei in Carter et al. (1999)
Reticulofenestra pseudoumbilicus-1	LO-1	n			“global”	3.820			Gi-10	Raffi & Flores (1995)
Globorotalia inflata-2	FO-2	f			Wairarapa (regional)	3.850	±	100	Gi-11	Shane et al. (1995)
Discoaster asymmetricus	LO	n	494.62	513.37	ODP 1119 (C52X-cc)	3.911		min. age	Gi-13	Wei in Carter et al. (1999)
Globorotalia puncticuloides-1	FO-1	f	494.62	513.37	ODP 1119 (C52X-cc)	3.911		min. age	Gi-13	Hayward in Carter et al. (1999)
Pseudoemiliana lacunosa-1	FO-1	n			“global”	4.000			Gi-16	Gartner (1977)
Globorotalia inflata (3 chambers)*-1*	FO-1	f	136.01	148.75	ODP 1123 (B16H-1-40)	4.135		+7.5	Gi-18	Scott (pers. comm.)
Base Thvera Subchron	reversal	M			(= base Opoitian)	5.235			TG-1	Lourens et al. in Gradstein et al. (2004)

Note: the listed first occurrences (FOs) of *Globorotalia oceania*-d include occurrences attributed in previous literature to *G. crassiformis*.

natural gamma ray time series for each segment of the figure. Historic onland stage and glacial terminologies are indicated as appropriate, and ancillary data relevant to climatic state, such as sand and carbonate content, are also provided. Table 1 comprises an important part of the database from which Fig. 7 was prepared, and is, of course, a work in progress. The first occurrences (FO) and last occurrences (LO) indicated in this table will be increasingly refined, and added to, by further research.

All correlation tools have errors (Carter in press), and potential correlation events are, therefore, shown on Fig. 7 and in Table 1 with their estimated error ranges. Pillans et al. (1996) have shown that in the Quaternary, oxygen isotope stratigraphy is superior to other practical numeric dating techniques, generally carrying an error of only a few thousand years back to 3 Ma and beyond. A similar resolution is achieved by magnetic reversals in tuned deep sea records (e.g., Tiedemann et al. 1994). Whereas the first or last occurrences of some planktic organisms such as *Stylatractus* may be able to match that resolution (e.g., Thierstein et al. 1977), it is rarely known in advance which index fossils are that accurate. New Zealand Middle-Late Cenozoic micropalaeontologic zonations have a working resolution at best of c. 1 m.y. (Cooper et al. 2001), and for benthic organisms, such as molluscs, the likely correlation error between different sections may range up to 2 m.y. or more. Furthermore, that benthic organisms are generally facies controlled, and sometimes specific to particular basins or geographic regions (e.g., Boreham 1963; Beu 1969; Neef 1970), further limits their usefulness in correlation.

Figure 7 and Table 1 demonstrate that possibly not a single Pliocene-Pleistocene index fossil has a synchronous FO or LO across the New Zealand region. A historic index, such as the FO of *Zygochlamys delicatula*, ranges from 2.47 Ma at Wanganui, 2.275 Ma at Mangaopari Stream, and 0.67 Ma at ODP Site 1119. Most other molluscs have comparable fluid ranges. Stenothermal warm water forms such as *Eunatacina*, *Anadara*, and *Capulus*—which, since Beu (2004), have well constrained ranges—are generally restricted to north-eastern (East Cape) or western (Wanganui) North Island, and have unknown or untested ranges countrywide. Planktonic microfossils fare little better, with the FO of *Globorotalia crassula* varying from 2.81 Ma at ODP Site 1119 to 2.44 Ma at Site 1123 to 2.40 in the Wairarapa regionally to 2.30 Ma at Mangaopari Stream; *Globorotalia puncticuloides* appears first at 3.91 Ma at Site 1119 and 3.10 Ma in the Wairarapa; *Globorotalia oceanica*—dextral (= *G. crassiformis* of earlier writers) is perhaps an exception, having its FO near 2.81 Ma at ODP Site 1119, at 2.91 Ma at Site 1123 and with overlapping error bars. Nannoplankton too have ranges that are obviously controlled by regional watermass, frontal and climatic factors, with the FO of *Emiliana huxleyi* ranging from 0.280 to 0.121 Ma; the reappearance of medium sized *Gephyrocapsa* ranging from 2.059 Ma at ODP Site 1119 to 1.730 at Site 1125; and the LO of *Pseudoemiliana lacunosa*, traditionally associated with the MIS 12/11 stage boundary at c. 0.42 Ma, varying between 0.68 Ma at Wanganui (near the MIS 17/16 boundary; Beu & Edwards 1984) and as young as 0.286 Ma (MIS 8) at ODP Site 1119.

Noting the ubiquity of these types of variations, inspection of Fig. 7 suggests two conclusions:

1. valuable though microfaunal and macrofaunal evidence is, other physical means of age determination have now become of paramount importance for furthering our understanding of New Zealand Pliocene-Pleistocene events; and
2. within the New Zealand region, correlation to the global MIS stage system will in future continue to be made using manifold criteria, which include lithology (cyclostratigraphy, tephrostratigraphy), numeric dating, magnetostratigraphy, chemostratigraphy, and biostratigraphy.

Few regions in the world offer such a well exposed and well studied Pliocene-Pleistocene history as does New Zealand, where the stratigraphic record also offers unparalleled opportunities to integrate climatic histories between atmosphere, land, and sea (e.g., Newnham et al. 1999). ODP Site 1119 suffers from the defect that its current age model lacks independent verification from, for example, palaeomagnetism or tephrochronology, and rests mainly on assumed cycle correlations with the Vostok ice core and oceanic oxygen isotope records. Further, the biostratigraphic age checks that are available for Site 1119 are prone to the limitations discussed above, though these limitations of course apply also to all other New Zealand successions. Therefore, despite its unconfirmed age model, the natural gamma ray record of Site 1119 serves as a valuable interim template against which to assemble other relevant stratigraphic data (Fig. 7). Together with the continuous or semi-continuous climatic histories drawn from other offshore drillsites (e.g., Nelson et al. 1985; Carter et al. 1999; Hall et al. 2001), the record from Site 1119 will serve well as a tool with which to study Pliocene-Pleistocene events in the New Zealand region.

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