

Magnetostratigraphic chronology of late Miocene to early Pliocene biostratigraphic and oceanographic events in New Zealand

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ABSTRACT

Successions of uplifted shallow marine sediments in New Zealand were among the first to provide evidence of latest Miocene climatic deterioration and were also among the first onshore successions where late Miocene to early Pliocene magnetostratigraphic records were established. A revised chronology of late Miocene to early Pliocene events has been determined from new magnetostratigraphic results from two important successions at Blind River and Upton Brook, South Island, New Zealand (lat. 41°45'S, long. 174°05'E).

Magnetostratigraphic results from Upton Brook are the first from strata containing the Kapitean index mollusks and provide the first reliable estimate of the age of the Tongaporuan-Kapitean boundary. A coherent chronology from Blind River and Upton Brook indicates that two synchronous low sea level and cool paleoclimatic events occurred in the late Miocene Kapitean Stage. This inference is supported from other well-known coeval successions in the New Zealand region. The revised New Zealand chronology supports recent determinations from the Mediterranean that indicate a two-stage "Messinian salinity crisis" and a 5.2-Ma age for the Miocene-Pliocene boundary. Data from New Zealand, the Mediterranean, and ocean cores around the world suggest that these latest Miocene events are related to glacio-eustatic sea level fluctuations. A 6.05-Ma age for the first occurrence datum of *Globorotalia conomiozea* indicates that significant diachroneity exists in the placement of this datum between the Southwest Pacific and the Mediterranean, confirming the observation that *G. conomiozea* is unsuitable for correlation between ocean basins.

INTRODUCTION

The latest Miocene was a time of major climatic and oceanographic change. Microfaunal assemblages (Kennett, 1967; Ingle, 1967; Kennett and Vella, 1975; Barron and Keller, 1983) and evidence for expansion of Antarctic glaciation (Shackleton and Kennett, 1975; Kennett, 1977) indicate that conspicuous mid- and high-latitude cooling occurred at that time. Other major events, during a relatively narrow interval from ca. 6.5–5 Ma, included worldwide marine regression (Kennett, 1967; Adams and others, 1977; Haq and others, 1987), isolation and desiccation of the Mediterranean basin and the "Messinian salinity crisis" (Hsü and others, 1973, 1977; Cita, 1982; Cita and McKenzie, 1986; Müller and Hsü, 1987), permanent decrease in oceanic carbon isotopic values (Loutit and Kennett, 1979; Keigwin, 1979; Keigwin and Shackleton, 1980; Haq and others, 1980; Hodell and others, 1986, 1989), an equatorward movement of the Antarctic convergence by ~300 km (Kennett, 1966a; Kennett and Watkins, 1974; Loutit, 1981), and a burst in evolutionary activity marked by the highest rate of faunal turnover for the whole Neogene (Wei and Kennett, 1986; Beu, 1990). Significant, synchronous events in the oxygen isotopic records from all of the world's oceans indicate two periods of glaciation in the latest Miocene that were followed by relatively warm interglacial periods (Hodell and Kennett, 1986; Hodell and others, 1986; Keigwin, 1987). These glacial events are estimated to have lowered sea level by ~40–60 m, contributing to the isolation of the Mediterranean basin during the Messinian (Hodell and Kennett, 1986). The presence of two evaporite units separated by a succession of marine marls in the Mediterranean basin suggests a two-stage Messinian salinity crisis, where the first period of desiccation was interrupted by an intra-Messinian inundation, followed by isolation and a

second period of desiccation (Hsü and others, 1977; Cita, 1982; McKenzie and Oberhaensli, 1985; Cita and McKenzie, 1986; Müller and Hsü, 1987). The Miocene epoch terminated abruptly when a marine transgression totally refilled the Mediterranean basin, ending evaporite deposition with full restoration of marine conditions (Cita, 1975, 1982; Cita and McKenzie, 1986). Early Pliocene transgressive sediments contain warmer faunal elements than terminal Miocene sediments, with concomitant decreases in $\delta^{18}\text{O}$ values indicating a marked glacial retreat (Cita and McKenzie, 1986; Hodell and others, 1986).

Successions of uplifted shallow marine sediments in New Zealand were among the first to provide evidence for latest Miocene climatic deterioration (for example, Kennett, 1967) and were also among the first onshore successions where late Miocene-early Pliocene magnetostratigraphic records were established. These pioneering studies established late Miocene-early Pliocene chronologies at Hinakura Road, Wairarapa (Lienert and others, 1972), Mangapoike River, Hawkes Bay, and Blind River in Marlborough (Kennett and Watkins, 1974; Fig. 1). The Blind River record is especially important because it is one of the first localities where combined biostratigraphic, magnetostratigraphic, and stable isotopic results were obtained (Kennett and Watkins, 1974; Loutit and Kennett, 1979).

From these early studies, the Miocene-Pliocene boundary was interpreted to lie near the middle of the Gilbert reversed Chron. These determinations caused an immediate international correlation problem, because the Miocene-Pliocene boundary stratotype, as defined in Italy, was considered to be significantly older (Berggren, 1973). An alternative interpretation of the Blind River magnetostratigraphy was published almost immediately by Ryan and others (1974), who aligned important faunal da-

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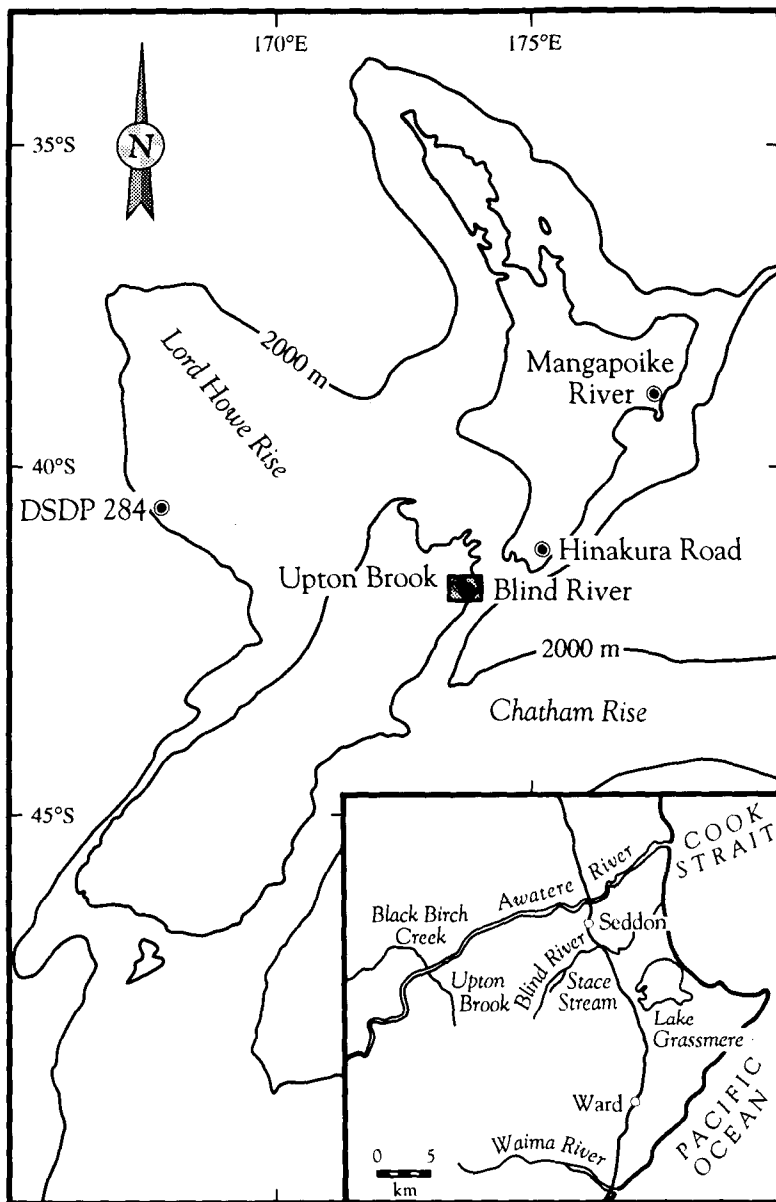


Figure 1. Map of New Zealand region showing locations where onshore late Miocene-early Pliocene magnetostratigraphic records have been obtained. Inset shows location of the Blind River and Upton Brook sections.

tum levels between Blind River and numerous Tasman Sea and Indian Ocean cores. This alignment placed the Miocene-Pliocene boundary near the bottom of the Gilbert reversed Chron, in closer agreement with the Mediterranean chronology, and resulted in closer alignment of Kapitean cold climatic intervals between Blind River and Tasman Sea cores.

Loutit and Kennett (1979) observed a "permanent" carbon isotopic shift in the middle part of the Blind River section (Fig. 2), which had previously been inter-

preted by Kennett and Watkins (1974) to be part of the Gilbert reversed Chron. A similar shift in $^{13}\text{C}/^{12}\text{C}$ is recorded in deep-sea cores from the Pacific and Indian Oceans within Chron 6 (Keigwin, 1979; Keigwin and Shackleton, 1980). Discovery of a shift in $\delta^{13}\text{C}$ at Blind River allowed Loutit and Kennett (1979) to make a definitive reinterpretation of the Kennett and Watkins (1974) record (Fig. 2), similar to that proposed by Ryan and others (1974).

In all of the early studies, spinner magnetometers were used to measure remanent

magnetization, and the demagnetization apparatus available were less efficient than modern equipment. Development of the cryogenic magnetometer has increased sensitivity by an order of magnitude in the measurement of remanent magnetization. As a result, recent work has shown that the early interpretations are not entirely correct. Wright and Vella (1988) extended the magnetostratigraphy at Mangapoike River to cover the whole of the late Miocene and demonstrated that the early work included some unreliable polarity determinations where their study overlapped with that of Kennett and Watkins (1974). Similarly, the magnetostratigraphy of Kennett and Watkins (1974) at Blind River has been shown to contain incorrect polarity determinations due to the incomplete removal of strong normal-polarity secondary overprints (Turner and others 1989).

The entire Blind River section studied by Kennett and Watkins (1974) was resampled during this study and that of Turner and others (1989). A parallel section at Upton Brook, 8 km west of Blind River (Fig. 1), was also sampled. A new magnetostratigraphic interpretation is presented here for both sections. The biostratigraphic implications of these chronologies are presented, and the constraints imposed by this study on the timing of paleoceanographic and paleoclimatic events in the terminal Miocene are discussed. The new results are consistent with global correlations and contribute to a higher resolution of terminal Miocene events. All dates cited are adjusted in accordance with the magnetic polarity timescale of Cande and Kent (1992).

BIOSTRATIGRAPHIC POSITION OF THE MIOCENE-PLIOCENE BOUNDARY IN NEW ZEALAND

A rapid evolutionary radiation of the *Globorotalia miotumida*¹ plexus is recognized during a short interval around the Miocene-Pliocene boundary (Hornibrook, 1984; Fig. 3). *G. miotumida* ranges from middle to upper Miocene (upper Lillburnian to mid-Kapitean) and is the dominant upper Miocene globorotalid in the New Zealand region (Scott and others, 1990). Sequential changes in the *G. miotumida* plexus were first studied

¹Numerous synonyms exist in the literature for most of the species under discussion. The names used here are those from the most recent revision of New Zealand foraminifera by Hornibrook and others (1989).

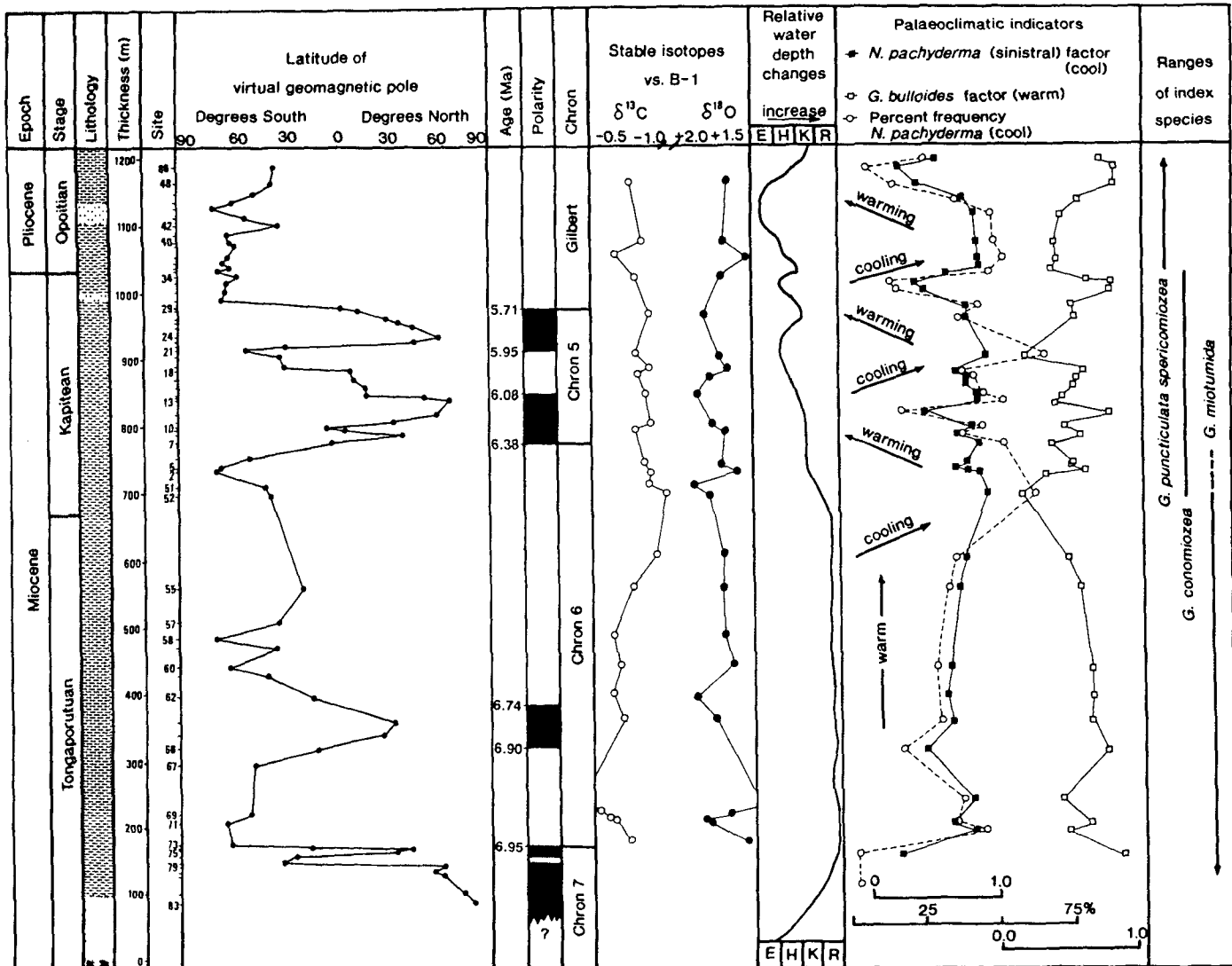


Figure 2. Stratigraphic succession, biostratigraphy, magnetostratigraphy, isotope stratigraphy, relative water depth changes, paleoclimatic history, and ranges of some key planktonic foraminiferal species at the Blind River section. Biofacies are represented by E (*Elphidium*), H (*Haeslerella*), K (*Karrerella*), and R (*Robulus*). After Loutit and Kennett (1979) and Loutit (1981), with ages of polarity boundaries adjusted to the magnetic polarity time scale of Cande and Kent (1992).

systematically by Kennett (1966a), who recognized a gradational change of *G. miotumida* to the more conical *G. conomiozea* in upper Tongaporutuan to Kapitean strata. Consistent recognition of the Tongaporutuan-Kapitean boundary in sequences lacking definitive mollusks is a difficult problem (Edwards and others, 1988; Hornibrook and others, 1989). Accurate location of this boundary at Blind River is currently not possible.

Separation of *G. conomiozea* as a distinct taxon from *G. miotumida* is also controversial. Scott (1979) does not recognize the replacement of *G. miotumida* by *G. conomiozea* at Blind River as shown by Kennett and

Watkins (1974) at site 52 (Fig. 2), but rather considers it to intergrade with crust-free forms of *G. miotumida*. Scott, (1979, 1980a) therefore regards *G. conomiozea* as an intraspecific conical variant of *G. miotumida* without separate taxonomic validity. Most New Zealand biostratigraphers, including Hornibrook and others (1989), recognize an interval with increasingly conical members of *G. miotumida* populations (irrespective of taxonomic status) and therefore retain the *G. conomiozea* zone in their biostratigraphic schemes.

The upper limit of *G. miotumida* or *G. conomiozea* is delineated by the appearance of *G. puncticulata sphericomiozea*, which

occurs over a short interval and is succeeded by *G. puncticulata puncticulata* (Fig. 3). *G. crassaformis* appears at a similar stratigraphic level as *G. puncticulata*; both forms are distinctive of the Pliocene. *G. inflata* becomes the dominant globorotalid in the upper Opoitian and is a major component of the planktonic foraminiferal assemblages of modern mid-latitude oceans (Scott, 1979; Hornibrook and others, 1989).

New Zealand biostratigraphers have consistently placed the Miocene-Pliocene boundary at the boundary between the Kapitean and Opoitian Stages (Finlay and Marwick, 1947; Jenkins, 1967; Collen and Vella, 1973; Kennett and Watkins, 1974).

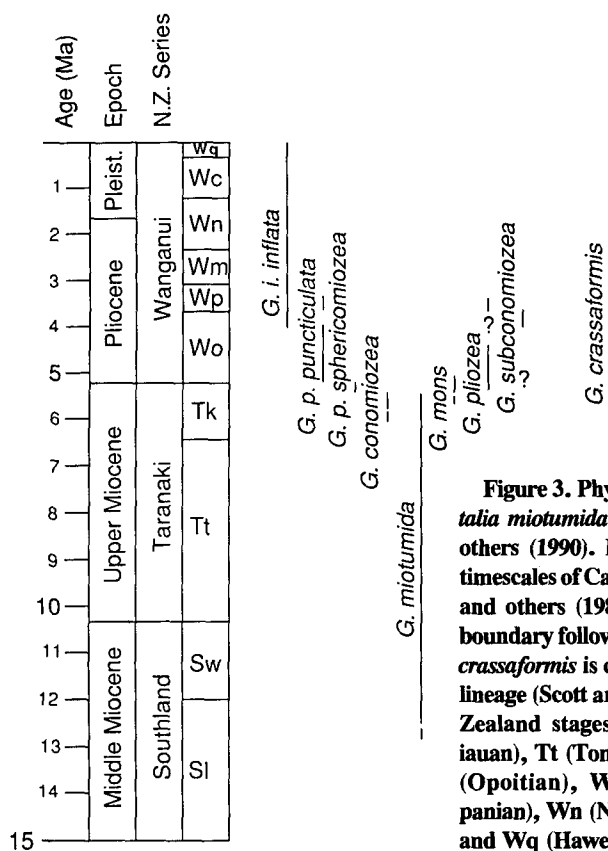


Figure 3. Phyletic arrangement of the *Globorotalia miotumida* lineage, adapted from Scott and others (1990). Dating of Miocene events follows timescales of Cande and Kent (1992) and Edwards and others (1988). Age of the Miocene-Pliocene boundary follows Hilgen and Langereis (1993). *G. crassaformis* is considered to be part of a separate lineage (Scott and others, 1990). Symbols for New Zealand stages are Sl (Lillburnian), Sw (Waiatauan), Tt (Tongaporutuan), Tk (Kapitean), Wo (Opoitian), Wp (Waipipian), Wm (Manganpanian), Wn (Nukumaruan), Wc (Castlecliffian), and Wq (Haweran).

Definition of the Kapitean-Opoitian boundary is problematical, however, because two different usages have developed during the last two decades (Scott, 1982). The first is that adopted by Kennett (for example, Kennett, 1966b; Kennett and Watkins, 1974; Hodell and Kennett, 1986), who places the Kapitean-Opoitian boundary at the first occurrence datum of *G. sphericomiozea*. Loutit and Kennett (1979) therefore placed the Miocene-Pliocene boundary at Blind River just above the base of the Gilbert reversed Chron at the first occurrence datum of *G. sphericomiozea*. In the second usage, the Miocene-Pliocene boundary is placed at the lowest joint occurrence of *G. puncticulata* and *G. crassaformis*. *G. sphericomiozea* defines a distinct upper Kapitean zone in this classification. This option is the preferred current usage in New Zealand foraminiferal biostratigraphy (for example, Collen and Vella, 1973; Scott, 1979; Hoskins, 1982; Edwards, 1987; Hornibrook and others, 1989) and is adopted here.

Correlation between the Miocene-Pliocene boundary in New Zealand and the type boundary in the Mediterranean has been restricted because of the provincial nature of New Zealand foraminifera and because no di-

rect paleomagnetic calibration of the Sicilian boundary stratotype has been available until recently. Langereis and others, (1984) obtained magnetobiostratigraphic results from a number of marine successions in Crete, which indicate that the second Messinian evaporite phase is younger than the base of the Gilbert reversed Chron. This interpretation was considered controversial by Berggren and others (1985) in their timescale, and they retained an age at the base of the Gilbert Chron for the Miocene-Pliocene boundary. Subsequently, Zijdeveld and others (1986), Hilgen and Langereis (1988), and Channell and others (1988) obtained magnetostratigraphic results from Sicily and Calabria in southern Italy. These data indicate that the Miocene-Pliocene boundary, defined by Cita (1975) at the base of the Trubi Marls, occurs within the Gilbert just below the Thvera Subchron. This age is consistent with $\delta^{18}\text{O}$ estimates that indicate the beginning of a long period of relatively warm paleoclimate just below the base of the Thvera Subchron (Hodell and Kennett, 1986; Hodell and others, 1986; Keigwin, 1987).

Hilgen and Langereis (1993) and Krijgsman and others (1993) reinterpreted the data of Langereis and others (1984), taking into

account the revisions to the magnetic polarity time scale by Cande and Kent (1992) who recognize two additional normal subchrons in Chron 6 which were not recognized in previous timescales (for example, Berggren and others, 1985). Consistent magnetobiostratigraphic dates have now been obtained in the Mediterranean for the Tortonian-Messinian boundary (Moreau and others, 1985; Channell and others, 1990; Benson and others, 1991; Benson and Rakic-El Bied, 1991; Krijgsman and others, 1993) and the Miocene-Pliocene boundary (Zijdeveld and others, 1986; Channell and others, 1988; Hilgen and Langereis, 1988, 1993). The increasing reliability of high-resolution studies of terminal Miocene events facilitate inter-regional and global correlation of the physical manifestations of these events. A 5.17-Ma age is assumed for the Miocene-Pliocene boundary in this study, following the conventional magnetic polarity time scale of Cande and Kent (1992) rather than the astronomically calibrated age of Hilgen (1991).

GEOLOGY OF THE BLIND RIVER AND UPTON BROOK SECTIONS

Blind River and Upton Brook (41°45'S, 174°05'E) expose a thick succession of siliciclastic marine sediments of the late Miocene to early Pliocene Awatere Group (Roberts and Wilson, 1992). The Awatere Group rests unconformably on indurated Jurassic basement rocks of the Torlesse Supergroup.

The lower 1200 m of the Blind River section was studied here and is relatively well-exposed at Blind River and Stace Stream. The basal Blind River deposits are ~100 m thick and consist of a thin basal conglomerate succeeded by medium sands with occasional shell beds (Roberts and Wilson, 1992). Silty mudstones dominate the remainder of the succession, with several thin sandstone beds occurring in Kapitean strata (Fig. 4).

Upton Brook lies 8 km west of Blind River and exposes a 1700-m succession that was deposited synchronously with that at Blind River, but in a shallower depositional environment (Roberts and Wilson, 1992). Distinct lithological units are recognizable in the Upton Brook section and are summarized in Fig. 4. Kennett (1966b) observed that despite overall facies differences between Upton Brook and Blind River, significant, and apparently synchronous, changes in depth of deposition are apparent in both sections, particularly in uppermost Tongaporutuan and lower Kapitean strata.

BIOSTRATIGRAPHY OF THE BLIND RIVER AND UPTON BROOK SECTIONS

Blind River

Biostratigraphic subdivision at Blind River by Kennett (1966b) and Kennett and Watkins (1974) is based on the evolutionary sequence of forms from *G. miotumida* to *G. puncticulata* (cited by Kennett, [1966a, 1966b] as *G. crassaformis*). Scott (1978, 1979, 1980a) took advantage of the thick succession at Blind River to make high-resolution studies of the phylogeny and biostratigraphy of *Bolivinita* and *Globorotalia* lineages. Kennett and Scott have different taxonomic concepts of some globorotalids and, as a result, have produced different biostratigraphic subdivisions, as outlined above. We follow Kennett (1966b) and Kennett and Watkins (1974) in recognizing *G. conomiozea* in our biostratigraphic scheme, but we follow the preferred usage in New Zealand biostratigraphy in placement of the Miocene-Pliocene boundary. This makes the Kapitean at Blind River thicker than recognized by Kennett and Watkins (1974; compare Figs. 2 and 4).

Planktonic foraminiferal populations indicate that the Blind River section was under sub-antarctic water throughout the entire late Miocene and earliest Pliocene (Kennett and Watkins, 1974). Populations of *Neogloboquadrina pachyderma* in present-day antarctic and arctic waters are consistently sinistrally coiled (Loutit, 1981). By analogy, prevalence of sinistral *N. pachyderma* in late Cenozoic sediments is taken to indicate cold water temperatures (for example, Ingle, 1967; Jenkins, 1967; Kennett, 1967). The paleoclimatic history determined at Blind River by Kennett and Watkins (1974), Loutit and Kennett (1979), and Loutit (1981), based on the relative frequency changes of *N. pachyderma* and *G. bulloides* (Fig. 2), indicate a change from relatively warm, stable conditions in the Tongaporutuan to greater surface water temperature oscillations during the Kapitean. Palynological assemblages from Blind River are characteristic of cool temperatures, while pollen from the Kapitean indicate a marked cooling relative to the rest of the sequence (Mildenhall, 1979). Benthonic foraminiferal biofacies indicate a shallowing of water depths during the latest Miocene (Kennett, 1966b; Fig. 2).

Upton Brook

The only detailed biostratigraphic analysis at Upton Brook is that of Kennett (1966b).

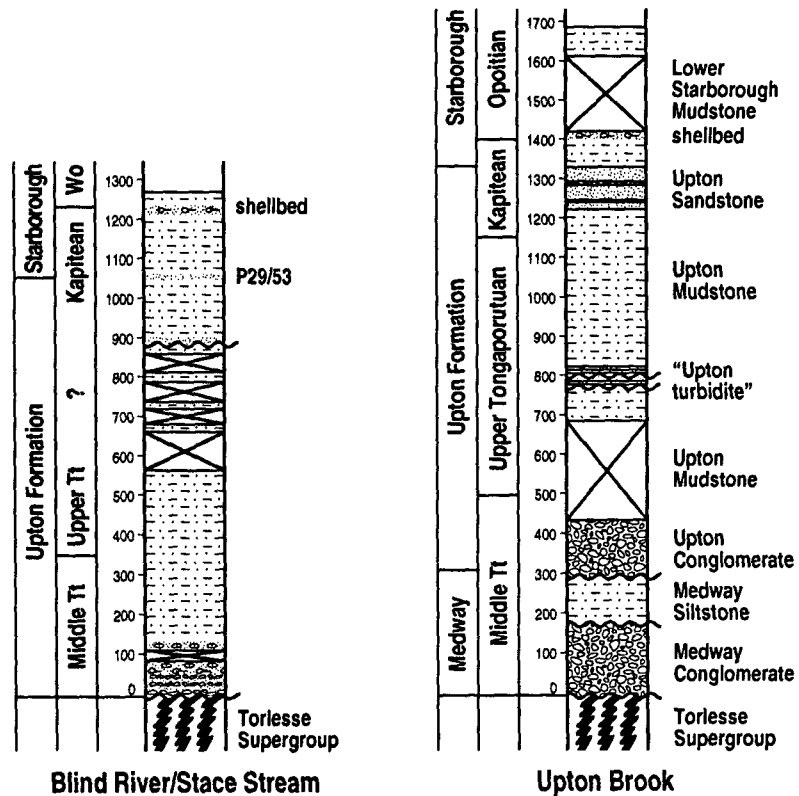


Figure 4. Stratigraphic succession, and the biostratigraphic and lithostratigraphic subdivision of the Blind River/Stace Stream and Upton Brook sections. Lithologic units and lithostratigraphic subdivision are after Roberts and Wilson (1992). Scale is in meters.

He distinguished Kapitean strata from the underlying Tongaporutuan and overlying Opoitian strata by the presence of key Kapitean mollusks. The base of the Kapitean stage is difficult to recognize at Blind River, because mollusks diagnostic of the Kapitean are restricted to relatively shallow-water facies, and Blind River sediments belong predominantly to a deeper-water facies (Kennett 1966b). The base of the Kapitean is recognized at Upton Brook at the first occurrence datum of *Sectipecten wollastoni* and *Austrofusus coeruleus* at 1,165 m (Fig. 4).

Shallow-water sediments at Upton Brook do not contain diagnostic planktonic foraminiferal species (Kennett, 1967); therefore, the top of the Kapitean is difficult to define. Kennett (1966b) placed the top of the Kapitean at a sandstone-to-siltstone contact at 1,330 m (Fig. 4), where a marked change occurs in the molluscan fauna along with a deepening of litho- and biofacies. This horizon occurs immediately above the highest occurrence of *Cucullaea*, which has its last occurrence in the Kapitean. Combined studies of planktonic foraminifera and mollusks indicate that all the Kapitean index mollusks, and the Mio-

cene lineages that became extinct in the Kapitean, did so at the end of the *G. conomiozea* zone (Beu and Maxwell, 1990). This event is by far the most marked extinction during the Neogene in New Zealand, when 23% of molluscan genera became extinct (Beu, 1990; Beu and Maxwell, 1990). The succeeding *G. sphericomiozea* zone contains an extremely impoverished fauna with far fewer taxa than any Pleistocene glacial fauna (Beu and Maxwell, 1990). In all sections studied by Beu and Maxwell (1990), including Blind River, the *G. sphericomiozea* zone is marked by deepening facies. *G. sphericomiozea* is not found at Upton Brook, but it is probable that the sequence correlative with the *G. sphericomiozea* zone starts at the marked change in litho- and biofacies at 1,330 m. This horizon is significantly below the Kapitean-Opoitian boundary as currently defined by New Zealand foraminiferal biostratigraphers. It is necessary, therefore, to realign the Kapitean-Opoitian boundary at Upton Brook to retain consistency with the Blind River biostratigraphy of Scott (1979, 1980a). The Kapitean-Opoitian boundary at Upton Brook is therefore placed at ~1,400

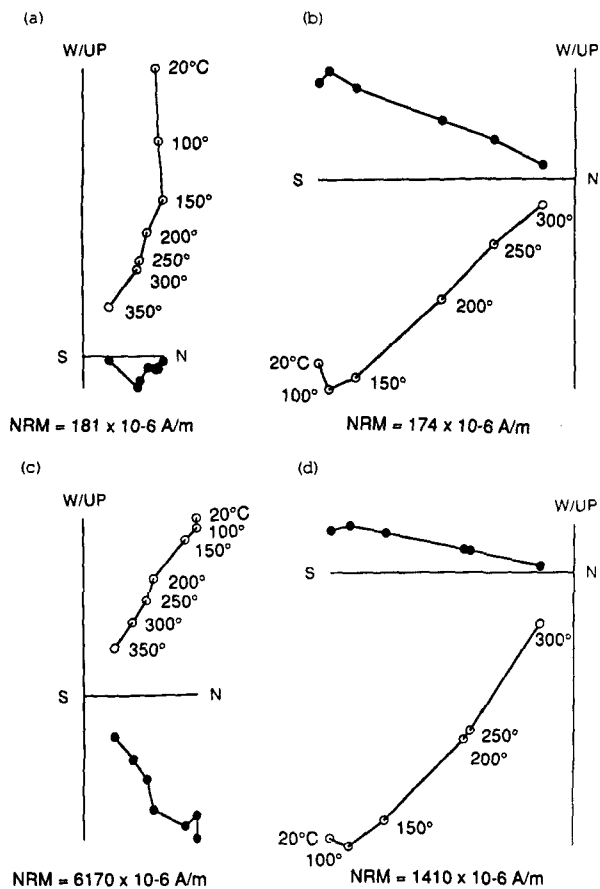


Figure 5. Vector component diagrams showing demagnetization behavior of samples from Blind River: (a) Normal polarity site 19; (b) Reversed polarity site 23. Upton Brook: (c) Normal polarity site 56; (d) Reversed polarity site 39. Solid circles denote projections onto the horizontal plane and open circles denote projections onto the vertical plane.

m, at a distinctive group of shell beds recognized at Blind River and Upton Brook (Beu and Maxwell, 1990; Roberts and Wilson, 1992).

METHODS

Paleomagnetic Sampling

Oriented paleomagnetic cores were sampled from fresh, unweathered, water-saturated outcrops. A minimum of three 25-mm-diameter cores were drilled from each site, with a maximum stratigraphic difference of 100 mm between the highest and lowest samples. Kennett and Watkins (1974) collected paleomagnetic samples from the lower 1,200 m of the Blind River section at Blind River and Stace Stream (Fig. 2). The lower 900 m of the section was sampled for this study, with 153 cores collected from 46 sites. Results from a detailed study by Turner and others (1989) are used here for the stratigraphic interval from 900 to 1,200 m.

No paleomagnetic work had been attempted at Upton Brook prior to this study. Samples were therefore collected from the

entire section, except for two covered intervals from 440 to 690 m and 1,410 to 1,630 m (Fig. 4). The confluence of Upton Brook with the Awatere River (Fig. 1) is at 1,410 m. The interval from 1,630 to 1,670 m was sampled at Black Birch Creek, which flows into Awatere River from the north (Fig. 1). A total of 274 cores were drilled from 80 sites. No sampling was possible at the base of the section, where thick conglomerates are unsuitable for paleomagnetic study.

Laboratory Procedure

All remanence measurements were made on a two-axis ScT cryogenic magnetometer at the Research School of Earth Sciences, Victoria University. Natural remanent magnetization was measured routinely on all samples to optimize the chances of selecting those most suitable for full stepwise demagnetization. Previous studies indicate that thermal demagnetization is superior to alternating field demagnetization in isolating primary remanence directions in New Zealand Cenozoic mudstones (Kennett and Watkins, 1974; Wright and Vella, 1988; Turner and oth-

ers, 1989; McGuire, 1989). Thermal demagnetization was therefore applied to the majority of samples demagnetized in this study, in a zero-field oven with temperature control within $\pm 5^\circ\text{C}$ of the desired temperature. Magnetic remanence and susceptibility were measured after heating to 100, 150, 200, 250, 300, 350, and 380 $^\circ\text{C}$. Magnetic susceptibility was measured using a Bartington Instruments MS1 magnetic susceptibility meter. Full stepwise thermal demagnetization was carried out on 232 samples from Blind River and 445 samples from Upton Brook. Large increases in susceptibility, usually of the order of 30% or more, accompanied heating between 300 and 380 $^\circ\text{C}$, marking the appearance of a magnetic phase displaying unstable and viscous magnetizations. Remanence measurements at temperatures where such dramatic changes in susceptibility occurred were not included in later analyses because any remaining primary magnetization is masked or obliterated.

Several strict data acceptance criteria were adopted. All samples were measured in four orientations, then inverted and remeasured in four orientations. Data were rejected if the angular difference between averaged directions calculated from the original and inverted positions exceeded 15° . Data also were rejected if the intensity of magnetization was less than $3 \times 10^{-5} \text{ Am}^{-1}$ or if magnetic susceptibility changed by 30% or more between demagnetization steps.

Magnetic Properties

Natural remanent magnetization intensities from both localities range from 10^{-5} to 10^{-1} Am^{-1} . Stable paleomagnetic behavior is observed in vector component plots from $\sim 50\%$ of the samples demagnetized (Fig. 5). The soft, low-temperature, normal-polarity, secondary component (removed below 250 $^\circ\text{C}$) is interpreted to be a Brunhes Chron viscous remanence. Remanence directions were determined from stably magnetized samples, such as those in Fig. 5, using a best-fit line that is constrained through the origin of the vector component plot. This best-fit remanence component is referred to as the characteristic magnetization.

For 35% of the samples demagnetized, the blocking temperature spectra of the characteristic and secondary components overlap to the extent that characteristic remanence directions are not isolated before thermal alteration occurs. In these cases, the trends described by measurements at low demagnetization levels are important for constraining

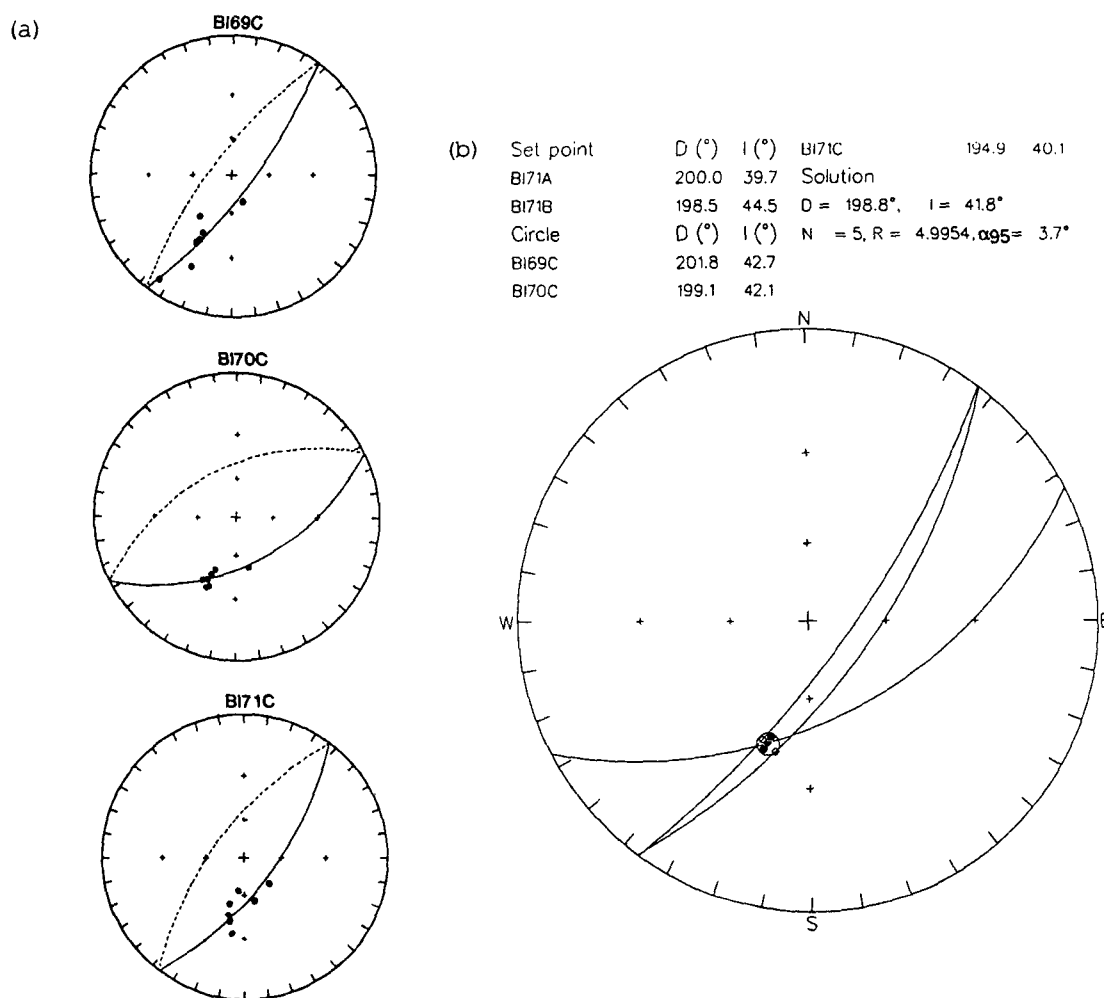


Figure 6. Example of use of the remagnetization circles analysis method of McFadden and McElhinny (1988) for obtaining the characteristic remanence direction for several samples from a site. Directions from stably magnetized samples (set points) are used to constrain the solution. Open circles denote upper hemisphere and solid circles denote lower hemisphere directions.

great circles of remagnetization (Fig. 6). The method of McFadden and McElhinny (1988) has been used extensively for sites where remagnetization great circles for numerous samples can be extrapolated to a convergence point which defines an "end point" or characteristic remanence direction for the site. Where possible, direct end point determinations from stably magnetized samples ("set points") are used to constrain such solutions (Fig. 6).

At some sites, there also appears to be a secondary component in the higher end of the blocking temperature spectrum. The high-temperature component is difficult to isolate due to thermal alteration and low intensities of magnetization. In such cases, the component residing in the middle of the blocking temperature spectrum is the most persistent and stable component and is interpreted to be

the characteristic component. It can be identified as straight lines that are not directed toward the origin of vector-component plots, and remanence directions were determined using a best-fit line that is not constrained through the origin of the vector-component plot. These directions are usually within a few degrees of those from single-component samples from nearby sites. The high blocking temperature component has been observed in other paleomagnetic studies of New Zealand mudstones (McGuire, 1989; Turner and Kamp, 1990) and may be a post-depositional chemical remanence residing in secondary minerals that have grown to stable, single domain size.

About 15% of samples from both localities display random, unsystematic behavior on demagnetization. Such samples are typically weakly magnetized ($< 1 \times 10^{-4}$ natural rem-

anent magnetization A_m^{-1}) and have normal-polarity secondary overprints. They are magnetically viscous and yield no useful paleomagnetic information. This behavior is probably due to low concentrations of ferrimagnetic material or the dominance of superparamagnetic and/or unstable magnetic phases.

Remanence Carriers and Timing of Remanence Acquisition

Rock magnetic and petrographic investigations indicate that the major contributors to remanence are ferrimagnetic iron-titanium oxide and iron sulfide (greigite and pyrrhotite) minerals (Roberts and Turner, 1993). Horizons where ferrimagnetic iron sulfide minerals are dominant are always extremely fine-grained and are usually intensely and sta-

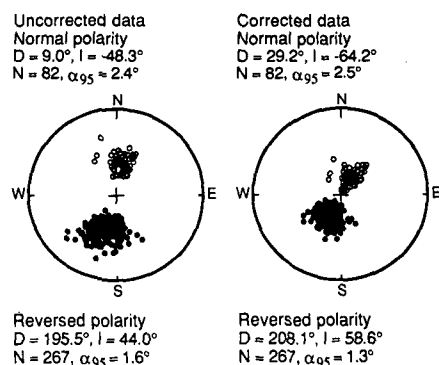


Figure 7. Equal angle stereographic projections of uncorrected and bedding-corrected characteristic remanence directions for combined data sets of stable samples from Blind River and Upton Brook. Open and solid circles denote upper and lower hemisphere projections, respectively.

bly magnetized. Such fine-grained horizons are stratigraphically restricted. Relatively coarser-grained siltstones dominate both successions, where the paleomagnetic signal is less intense and generally less stable than in fine-grained horizons. Titanomagnetite and maghemite are the only ferrimagnetic minerals identified in the coarser-grained horizons (Roberts and Turner, 1993).

Post-depositional remanent magnetization is a common remanence acquisition mechanism in many deep and marginal seas where bioturbation is almost ubiquitous (Verosub, 1977). Lack of an inclination error in the extensively bioturbated sediments from Blind River and Upton Brook (Fig. 7) provides evidence for a post-depositional acquisition of remanence. Iron-titanium oxide grains in the Awatere Group have a detrital origin (Roberts and Turner, 1993) and would be expected to give rise to a post-depositional remanent magnetization that can be "locked in" up to several thousand years after deposition (for example, deMenocal and others, 1990).

Available evidence suggests that sedimentary iron sulfide formation is severely restricted below the surface meter of marine sediments (Westrich and Berner, 1984; Canfield and Berner, 1987). If so, ferrimagnetic iron sulfides will form during early diagenesis, within 2,000 yr of deposition for sediments deposited at rates >0.5 m/1,000 yr. Special cases exist where dissolved sulfate, which reacts with ferrous ions to form iron sulfide, is available at greater depths, giving rise to iron sulfides during later diagenesis. In these cases, however, sulfate is derived from deeper evaporite deposits (Channell and

Hawthorne, 1990) and from hydrocarbon seepage (Reynolds and others, 1990). Such conditions do not exist in the Awatere Group; thus, the most likely time for sulfide formation is during burial at depths <1 m, where dissolved sulfate is available to react with soluble iron compounds (Roberts and Turner, 1993).

Normal and reversed polarity data (Fig. 7) pass the paleomagnetic reversal test of McFadden and McElhinny (1990) and correction for tectonic tilt results in inclinations as expected at the site latitude for an axial geocentric dipole field ($I_{exp} = 60.7^\circ$). Clockwise rotation from the expected axial dipole declination is due to post-late Pliocene tectonic rotation (Roberts, 1992). A late Pliocene angular unconformity in sediments of the Awatere Group suggests that this deformation began at ca. 3 Ma (Roberts and Wilson, 1992; Roberts, 1992). The reversal and fold tests indicate that the characteristic remanence directions shown in Figure 7 are relatively free of secondary contaminations and predate the deformation that tilted the strata to their present attitude (that is, the magnetization is >3 Ma). Furthermore, the dominance of reversed polarity over normal polarity directions (Figs. 7 and 8) is consistent with a late Miocene acquisition of magnetization, because reversed polarity is dominant from Chron 6 to the early Gilbert.

Comparison of paleomagnetic directions from fine-grained sulfide-bearing sites with those from immediately over- or underlying coarser-grained iron-oxide-dominated sites indicates that there are no significant differences in paleodirection. This, and other factors discussed above, suggests that both remanence carriers had their remanence locked in within a few thousand years of deposition. Such a lag between deposition and acquisition of magnetization is insignificant compared to the resolution of the magnetic polarity time scale. This conclusion is compatible with other magnetostratigraphic studies that demonstrate that sulfide minerals can be responsible for the stable characteristic remanences that pass various paleomagnetic tests indicating acquisition of remanence in the early stages after deposition (Linszen, 1988; Tric and others, 1991; Van Hoof and Langereis, 1992).

RESULTS

The Blind River Record

Detailed sampling (308 sites) over the interval from 900 to 1,200 m by Turner and oth-

ers (1989) was aimed at obtaining records of geomagnetic polarity transitions in Chron 5, based on the results of Kennett and Watkins (1974; Fig. 2). Turner and others (1989) found that Kennett and Watkins (1974) failed to recognize a normal polarity overprint on a reversed polarity primary component in the uppermost interval of normal polarity (Figs. 2 and 9). An example of the type of paleomagnetic behavior observed by Kennett and Watkins (1974) at sites 23–29 is also seen in our data obtained near site K&W14 at Stace Stream (ST3; Fig. 9). Natural remanent magnetization data are relatively well-grouped with normal polarity directions ($D = 19.0^\circ$, $I = -77.2^\circ$, $\kappa = 7.4$, $\alpha_{95} = 20.2^\circ$, $N = 9$). Scatter away from normal and toward reversed polarity occurs on demagnetization, without isolation of a consistent reversed primary direction. It is possible to fit great-circle paths to data from four thermally demagnetized samples (Fig. 10), although the great-circle fits are inaccurate due to the low signal/noise ratio. Despite scatter in the data, demagnetization paths can be extrapolated to a coherent, reversed polarity end-point direction ($D = 216.4^\circ$, $I = 72.7^\circ$, $\kappa = 79$, $\alpha_{95} = 20.1^\circ$). Thermal demagnetization of nine samples collected from site 24 of Kennett and Watkins (1974) in Stace Stream also show well-clustered, stable, reversed polarity directions ($D = 211.8^\circ$, $I = 60.1^\circ$, $\kappa = 353$, $\alpha_{95} = 4.2^\circ$) in direct contradiction to the interpretation of Kennett and Watkins (1974).

Given the importance of the Blind River record, the remainder of the section was sampled to check the reliability of the Kennett and Watkins (1974) record. We have used the stratigraphic column of Morgans (1980) and the revised biostratigraphy of Scott (1979; Fig. 9).

New magnetostratigraphic results from the lower Blind River section (Fig. 9) differ little from those of Kennett and Watkins (1974). Reversed polarity directions dominate the lower 560 m of the succession, where only two intervals of normal polarity are recognized (Figs. 8 and 9). Coarse-grained rocks, which are unsuitable for paleomagnetic study, dominate the lower 135 m of the succession, and we are unable to reproduce the results of Kennett and Watkins (1974) in that interval. No significance should be attached to their normal-polarity results from this interval, because no demagnetization was performed on samples from site K&W 79 and below. We conclude that the polarity below 135 m is indeterminate. Documentation of the Chron 6 $\delta^{13}\text{C}$ shift in upper Tongaporutuan strata at Blind River and the general repro-

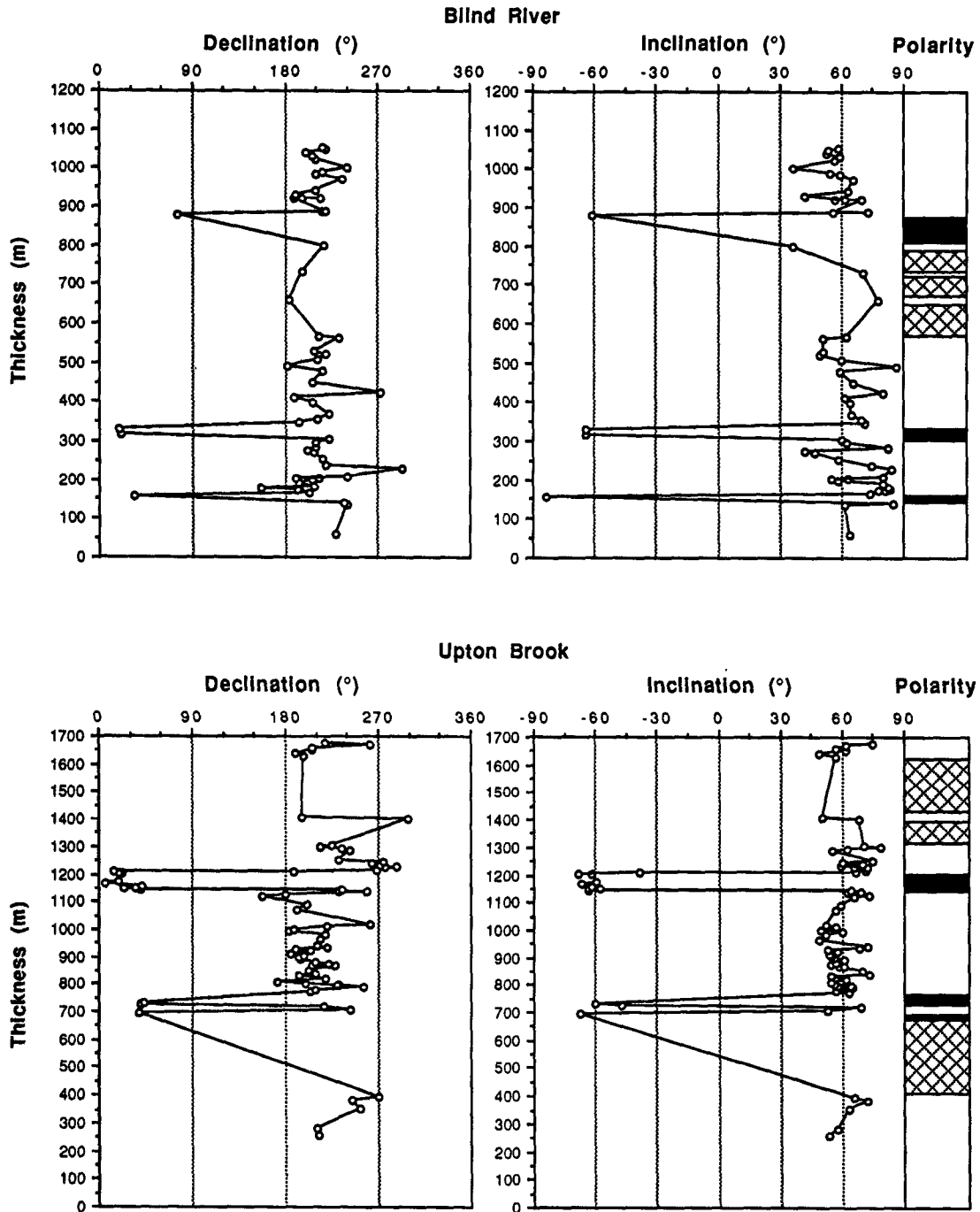


Figure 8. Site-averaged paleomagnetic directions versus stratigraphic position for Blind River and Upton Brook. Data from Turner and others (1989) are included in the upper Blind River record. All directions from the lower Blind River and Upton Brook records are from fully stepwise-demagnetized samples. Open circles denote sites with single-component magnetizations, solid circles denote sites where remagnetization circles analysis was used. Black and white intervals on the polarity log denote normal and reversed geomagnetic polarity, respectively, and cross-hatched intervals indicate that exposure is either covered or inaccessible.

ducibility of the Kennett and Watkins (1974) magnetostratigraphy in the lower Blind River section add confidence to the stratigraphic interpretation of Loutit and Kennett (1979).

No evidence has been found in our studies to contradict the existence of the zone of nor-

mal polarity recorded by Kennett and Watkins (1974) in the poorly exposed interval from 812 to 888 m. Reversed polarity data are recorded from sites ST3 (Stace Stream) and BB5 (Blind River) at the top of this interval, at 888 m, as well as from site ST2 (Stace

Stream) at the base of this interval, at 800 m (Fig. 9). The only truly normal polarity results obtained in our studies in the interval from 812 to 888 m come from a site just below a thin turbidite sequence that is recognized at the same stratigraphic position (880 m) at

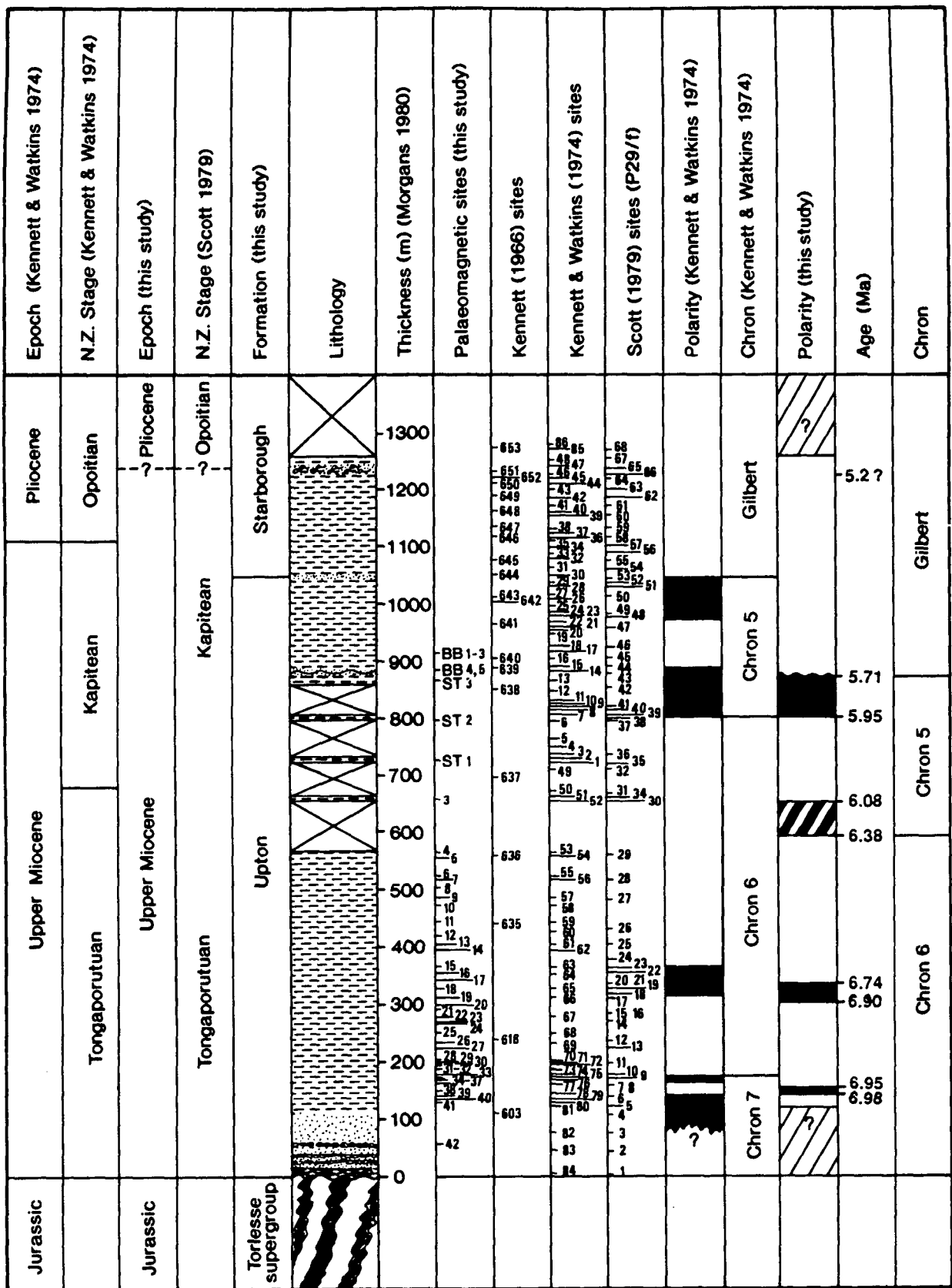


Figure 9. Composite stratigraphic succession at Blind River and Stace Stream, showing biostratigraphy, sampling sites, and revised and previous magnetostratigraphy on the thickness scale of Morgans (1980). Solid diagonal bars on polarity log indicate inferred normal polarity; thin diagonal bars indicate unknown polarity. The sites of Turner and others (1989) are not shown.

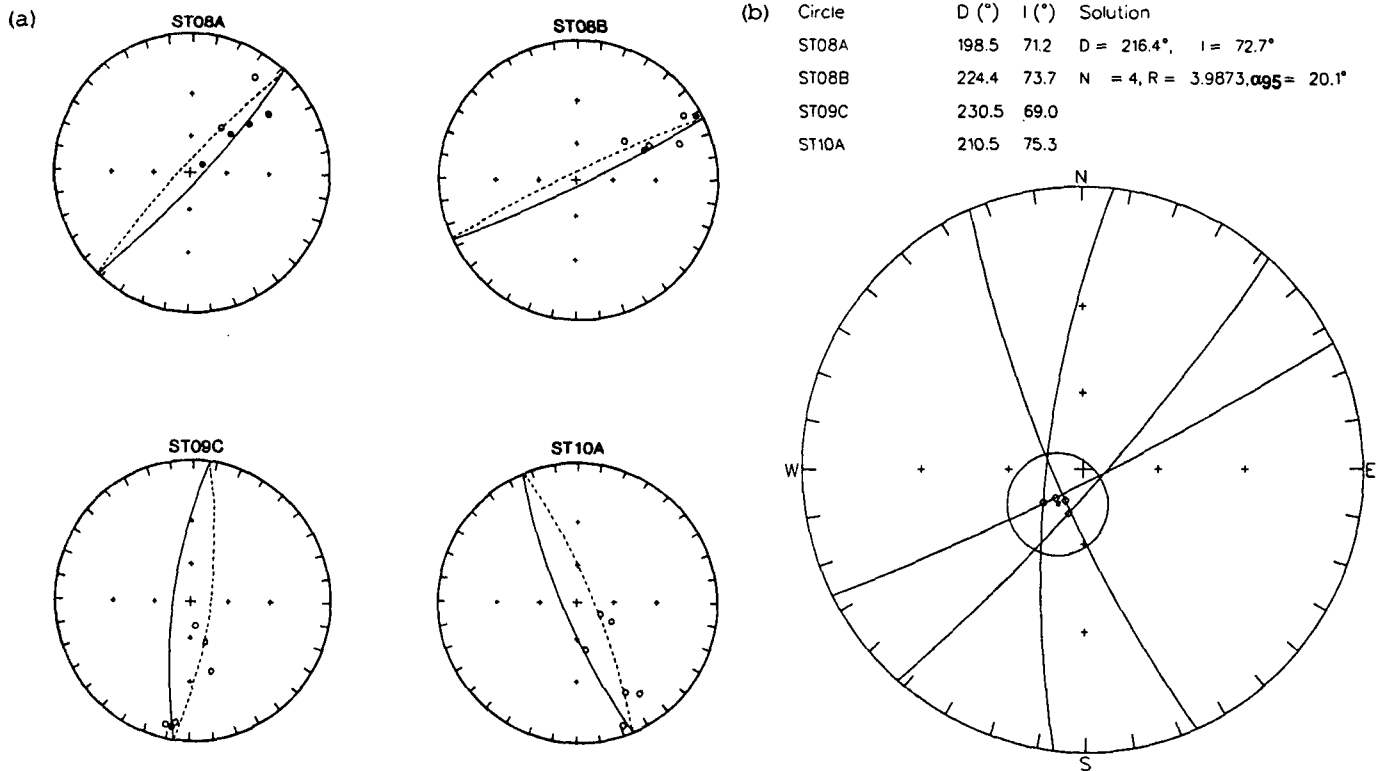


Figure 10. Remagnetization circles analysis of strongly overprinted samples from Stace Stream site ST3: (a) Great circle fits to thermal demagnetization data; (b) Extrapolation of great circles to convergence at a reversed polarity end point. Symbols as in Figure 6.

Stace Stream, Blind River, and Cecilia Creek (Roberts and Wilson, 1992). Alternating field demagnetization isolated a relatively close grouping of normal polarity directions ($D = 66.6^\circ$, $I = -58.5^\circ$, $\kappa = 23$, $\alpha_{95} = 10.3^\circ$, $N = 10$), at a site located between sites 13 and 14 of Kennett and Watkins (1974). The upper part of this normal subchron is interpreted to have been truncated by the turbidite unit, because it occurs between two sites of opposite geomagnetic polarity, although no significant time break is evident there. The base of this normal subchron is inferred to occur in the interval between 800 and 835 m (K&W sites 6–11), where Turner and others (1989) obtained predominantly intermediate polarity directions.

Turner and others (1989) suggested that if the two normal subchrons of Chron 5 exist at Blind River, then both must be lower than reported by Kennett and Watkins (1974), because the uppermost reversed portion of the record is substantially thicker than reported previously. The long, reversed polarity interval is interpreted to correspond to the base of the Gilbert reversed Chron. Biostratigraphic resolution in the late Miocene at Blind River is sufficient to discount the possibility that the

turbidite sequence at 880 m has removed one of the normal subchrons of Chron 5, particularly because this interpretation requires a hiatus of at least 200,000 yr. The normal interval between ~812 and 880 m is therefore interpreted to correspond to the upper Chron 5 normal subchron. The lower Chron 5 normal subchron is inferred to lie in an interval of covered section between 560 and 800 m. Reversed polarity data are recorded from site ST1 at Stace Stream (733 m) and site 3 at Blind River (660 m; Fig. 9). The normal subchron, therefore, probably lies in one of three unexposed intervals from 560 to 660 m, 660 to 733 m, or 733 to 800 m.

Lack of exposure in a crucial part of the Blind River record seriously compromises its usefulness. Upton Brook was selected for magnetostratigraphic study because of good biostratigraphic control of upper Tongaporuan and Kapitean strata.

The Upton Brook Record

Reversed polarity characteristic remanence directions dominate the Upton Brook record. Three zones of normal polarity are recognized (Fig. 11). The lowermost occurs

at the top of an interval of covered section and is represented only by site 60 (694 m). A broader zone of normal polarity is recorded at sites 57 and 56 (727–732 m). Site 55 represents the start of a thick zone of reversed polarity from 750 to 1,140 m, which begins below the base of a thick turbidite unit at 760 m. A thick zone of normal polarity extends between sites 18 and 7 (1,146–1,208 m).

All sites sampled above 1,210 m at Upton Brook and Black Birch Creek display reversed polarity directions. The interval from 1,410 to 1,630 m was not sampled, because the Awatere River makes exposure inaccessible (Fig. 1). The relatively long interval of reversed polarity at the top of the Upton Brook record can be correlated with that at the top of the Blind River record and is interpreted to represent the base of the Gilbert reversed Chron. It is possible that the lowermost normal subchron of the Gilbert (Thvera) occurs in the unsampled interval (1,410–1,630 m).

The base of the normal polarity zone from 1,146 to 1,208 m is interpreted to represent the base of Chron 5, which could not be sampled at Blind River because of lack of exposure. The upper normal subchron of Chron 5

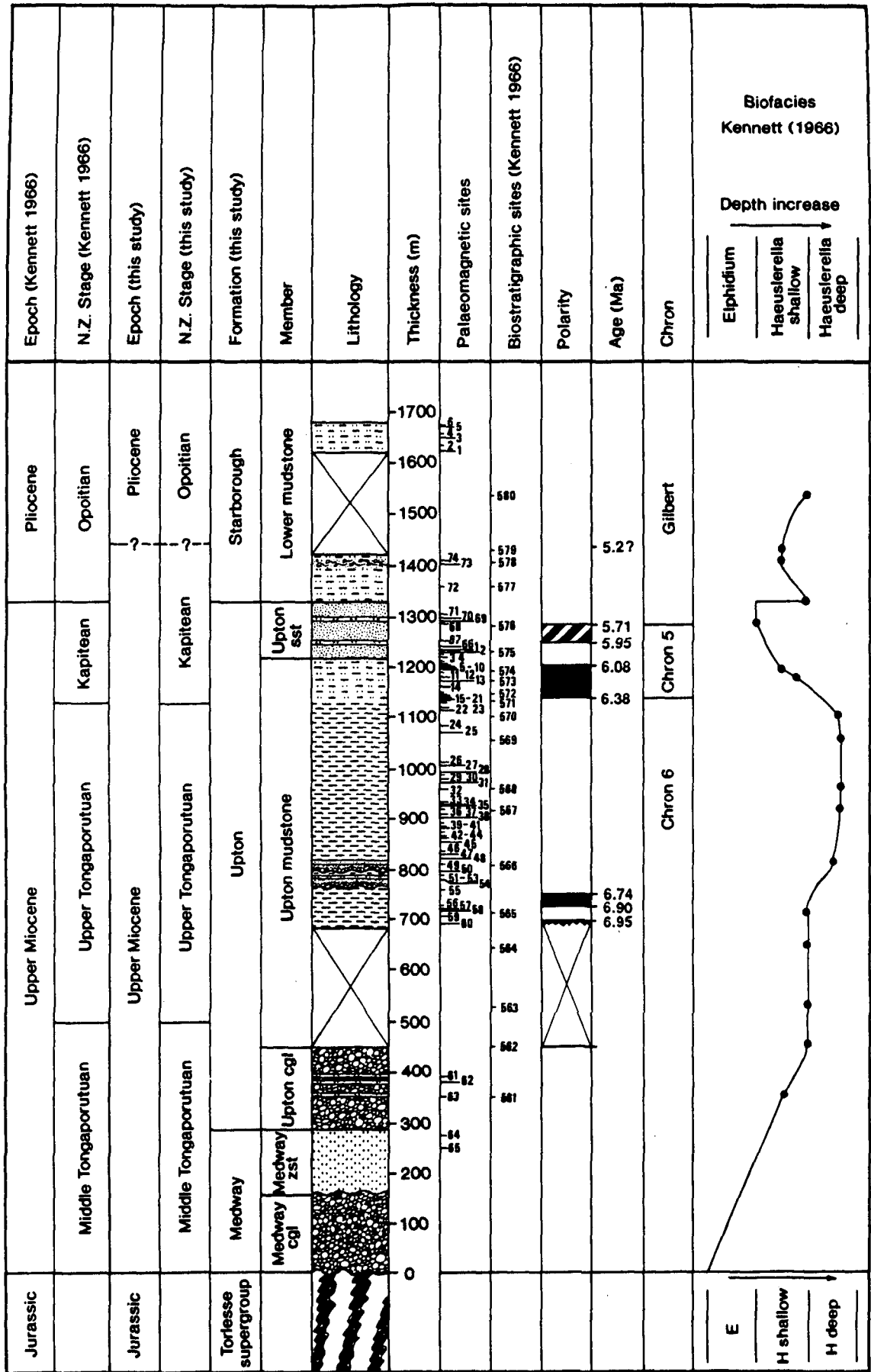


Figure 11. Stratigraphic succession at Upton Brook showing biostratigraphy, magnetostratigraphy, sampling sites, and relative water-depth changes. Solid diagonal bars on the polarity log indicate inferred normal polarity.



has not been recorded, either, because it occurs within the Upton Sandstone (between sites 67 and 68 [1,251–1,288 m]; Fig. 11), which is unsuitable for paleomagnetic sampling, or because it is missing from the record because of erosion. The most likely position for an erosional break is the sandstone-to-siltstone contact that separates the Upton and Starborough Formations at 1,330 m. If the upper normal subchron of Chron 5 has been eroded from the record at this position, a sedimentary hiatus of ~200,000 yr is required.

The long interval of reversed polarity from 1,140 to 750 m is inferred to represent reversed polarity in Chron 6. Three zones of normal polarity occur in Chron 6 in the magnetic polarity time scale of Cande and Kent (1992). A thick sequence of turbidites occurs from 760 to 820 m, directly above two of the Chron 6 normal subchrons. Their presence can account for higher rates of sedimentation in this part of the sequence. Wright and Vella (1988) record turbidite beds, which removed the Chron 6 normal subchrons, at Mangapoike River, Hawkes Bay. The turbidite sequence in the Upton Formation is recognized only in successions deposited at relatively shallow water depths (for example, Upton Brook) and does not have correlatives in deeper water successions such as at Blind River (Roberts and Wilson, 1992). We infer that the lowest normal subchron of Chron 6 lies in the interval of covered section below 700 m.

Few sites could be sampled below site 60 because of poor exposure and thick conglomeratic sequences that are unsuitable for paleomagnetic sampling. Reversed polarity directions obtained from thin silt horizons within the Upton Conglomerate (sites 61–63) and Medway Siltstone (sites 61–65; Fig. 11) cannot be confidently correlated to the magnetic polarity time scale.

A Composite Magnetostratigraphic Record

The Blind River and Upton Brook sections both lack exposure in critical parts of the

record. Sufficient lithostratigraphic and biostratigraphic control exists, however, (Roberts and Wilson, 1992) to allow construction of a composite record (Fig. 12). Both sections have long zones of reversed polarity at the top of the record, which are correlated with the lower Gilbert reversed Chron. The shallower-water Upton Brook record has coarse-grained sediments that are not suitable for paleomagnetic analysis in the upper Kapitean, while the Blind River record is unexposed for most of the lower Kapitean. However, combination of the two data sets results in a composite magnetostratigraphy which enables a one-to-one correlation with the MPTS of Cande and Kent (1992) from the lower Gilbert to lower Chron 6 (Fig. 12).

AGES OF BIOSTRATIGRAPHIC DATUM PLANES

Revision of the Blind River chronology and new constraints imposed by the Upton Brook magnetostratigraphy require modifications to previous correlations between magnetostratigraphically dated onshore New Zealand sedimentary successions. A chronostratigraphic column for late Miocene and early Pliocene time, revised from those of Edwards (1987) and Wright and Vella (1988), is presented in Fig. 13.

Chron 6 $\delta^{13}\text{C}$ Shift

The Chron 6 carbon shift is a globally recognized datum, with isotopically lightest values occurring ~100,000 yr after the upper normal subchron in Chron 6 (Vincent and others, 1980, 1985; Hodell and others, 1986). Inadequate outcrop in the vicinity of the $\delta^{13}\text{C}$ shift at Blind River prevented Loutit and Kennett (1979) from precisely determining its position (Fig. 2). The lightest $\delta^{13}\text{C}$ values recorded at Blind River occur between 580 and 660 m on the thickness scale of Fig. 9, which would place the $\delta^{13}\text{C}$ "shift" (of ~0.5‰) within Chron 5 in the new magnetostratigraphic interpretation. However, some carbon isotope records from the deep sea have $\delta^{13}\text{C}$ "shifts" of up to 1‰ within the upper Miocene, above the Chron 6 $\delta^{13}\text{C}$ shift (for example, Vincent and others, 1985). We believe that the lack of stratigraphic control on the position of the $\delta^{13}\text{C}$ shift, due to inadequate outcrop and insufficient sampling resolution, prevents the observed carbon shift from being a useful point for detailed correlation in the Blind River record. Given the above uncertainty and that few points were sampled in Chron 6, we cannot be sure that

the recorded carbon shift is actually the so-called "Chron 6 $\delta^{13}\text{C}$ shift." We interpret that the actual Chron 6 $\delta^{13}\text{C}$ shift has not been detected due to low sample density.

Planktonic Foraminifera

First Occurrence Datum of *Globorotalia conomiozea*. Kennett and Watkins (1974) recognized an evolutionary change from *G. mitotumida* to *G. conomiozea* at 660 m (P29/f30) at Blind River. This datum is inferred to occur near the base of the reversed part of Chron 5 in the revised magnetostratigraphy presented above (Figs. 9 and 13) at ca. 6.05 Ma as opposed to the top of Chron 6 (ca. 6.5 Ma) as reported by Loutit and Kennett (1979). This revision is in agreement with the first occurrence datum of *G. conomiozea* at Deep Sea Drilling Project (DSDP) Site 588 (South Pacific) which Hodell and Kennett (1986) report in the reversed part of Chron 5. This age is also in agreement with the interpretation of Edwards (1987) of DSDP Site 284 and the assertion by Wright and Vella (1988) that most of the *G. conomiozea* zone at Mangapoike River occurs within a diastem that includes all of Chron 5.

We reported a possible convergence of ages of the first occurrence datum of *G. conomiozea* (Turner and others, 1989) based on the above and the findings of Langereis and others (1984). Despite this apparent agreement, Edwards (1987) reports a discrepancy of up to 0.9 m.y. in the placement of this datum in the Southwest Pacific. New work on the Cretan sections by Krijgsman and others (1993) supercedes that of Langereis and others (1984) and has led to a magnetostratigraphic calibration in which *G. conomiozea* first appears consistently in a zone of reversed polarity at 6.92 Ma. This datum is recorded in normal polarity strata at ca. 6.8 Ma in the Tyrrhenian Sea at Ocean Drilling Program Site 654 (Channell and others, 1990) and on the Atlantic margin of Morocco (Moreau and others, 1985). The first occurrence datum of *G. conomiozea* is recorded slightly higher at the boundary between normal and reversed polarity at ca. 6.75 Ma in other Moroccan sections (Hodell and others, 1989; Benson and others, 1991; Benson and Rakic-El Bied, 1991).

As the age of the first occurrence datum of *G. conomiozea* is becoming better constrained, it appears that significant diachroneity exists in the placement of this datum between the Mediterranean and the Southwest Pacific. Discordance in this datum between Morocco and the Southwest Pacific

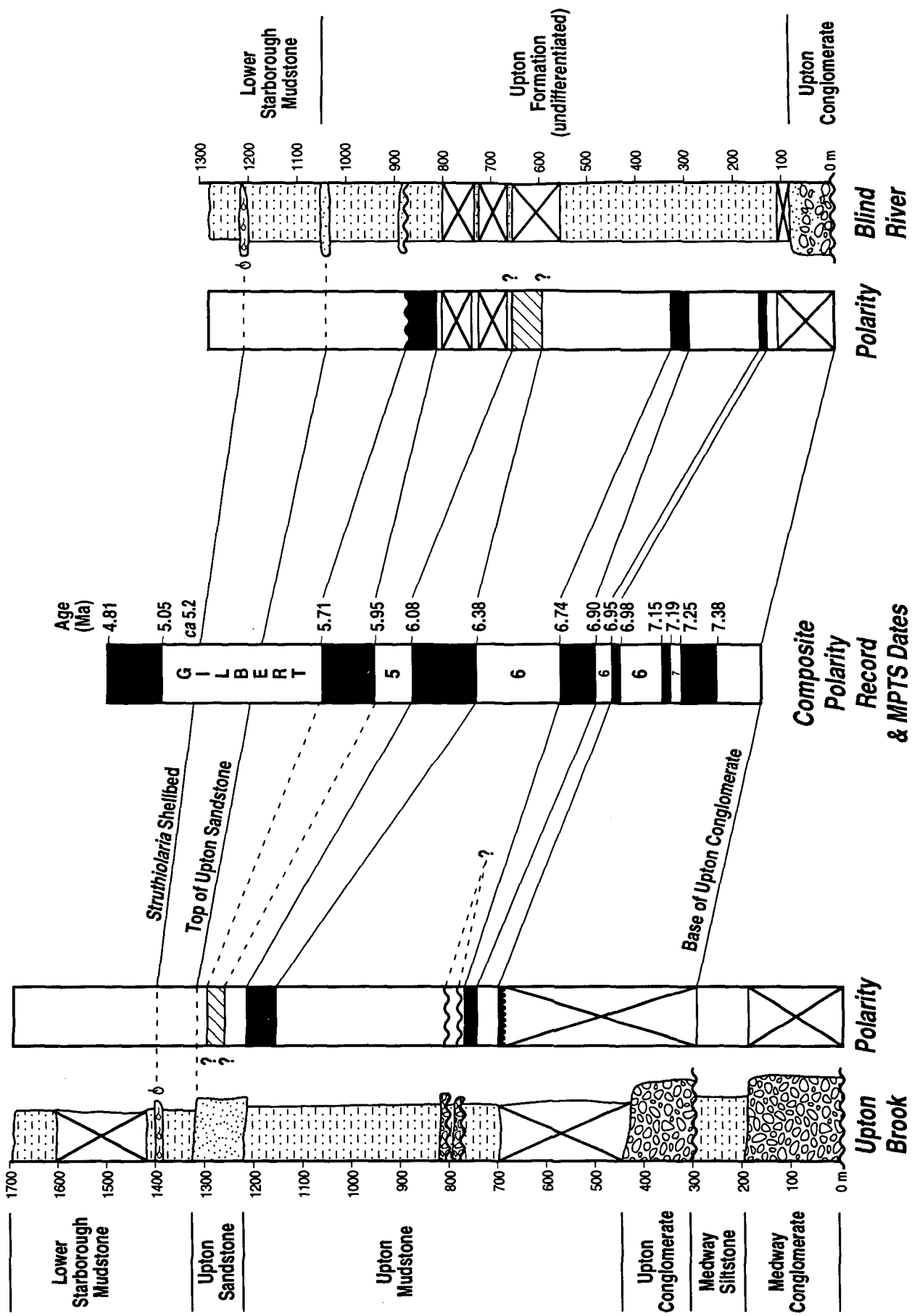


Figure 12. Lithostratigraphy and composite magnetostratigraphy for the Blind River and Upton Brook sections, and correlation with the magnetic polarity time scale (MPTS) of Cande and Kent (1992). Constraints on lithostratigraphic and biostratigraphic correlations are from Roberts and Wilson (1992).

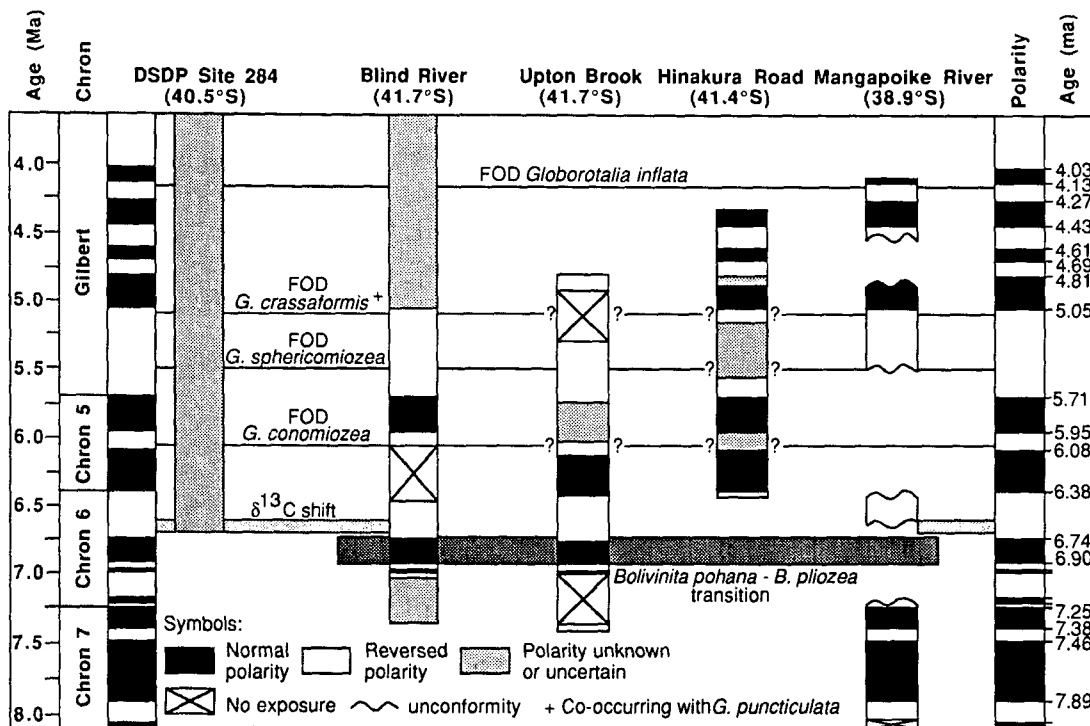


Figure 13. Late Miocene-early Pliocene chronostratigraphic correlation of onshore New Zealand successions with biostratigraphy and stable isotope stratigraphy of Deep Sea Drilling Project Site 284, adapted after Edwards (1987) and Wright and Vella (1988), using constraints imposed by this study and the magnetic polarity time scale of Cande and Kent (1992).

led Hodell and Kennett (1986) to suggest that the first occurrence datum of *G. conomiozea* is unsuitable for correlation between ocean basins. Furthermore, Scott (1980b) claims that significant morphological differences exist between specimens identified as *G. conomiozea* from New Zealand and Mediterranean strata and concluded that the Pacific and Mediterranean taxa may not be related.

First Occurrence Datum of *Globorotalia sphericomiozea*. The magnetostratigraphic interpretation of Loutit and Kennett (1979) placed the first occurrence datum of *G. sphericomiozea* within the lower reversed portion of the Gilbert Chron. The lower reversed portion of the Gilbert is ~170 m thicker in the revised magnetostratigraphy presented above. Extrapolating the average sedimentation rate of 0.55 m/1,000 yr from the top of Chron 5 at 880 m to the first occurrence datum of *G. sphericomiozea* at 1,080 m (P29/f55, Scott (1979)) gives a revised age of ca. 5.3 Ma for this datum. Dating of the first occurrence datum of *G. sphericomiozea* in the New Zealand region is largely dependent on the magnetostratigraphy at Blind River, because this datum is not recognized at Mangaopari Stream or Hinakura Road

(Edwards, 1987). At Mangapoike River, this datum occurs within a sedimentary hiatus that represents more than 0.5 m.y. (Wright and Vella, 1988) and, therefore, cannot be dated accurately. The first occurrence datum of *G. sphericomiozea* corresponds closely with the highest known occurrences of Kapitean mollusks (Scott, 1982; Beu and Maxwell, 1990). An age of ca. 5.5 Ma is obtained for the last occurrence datum of Kapitean mollusks at Upton Brook by extrapolating Kapitean sedimentation rates to 1,330 m. An age of 5.3–5.5 Ma is therefore suggested for the first occurrence datum of *G. sphericomiozea* in the New Zealand region.

Joint First Occurrence Datum of *Globorotalia puncticulata* and *G. crassaformis*. The lowest joint occurrence of *G. puncticulata* and *G. crassaformis* is used to identify the Kapitean-Opoitian and Miocene-Pliocene boundaries in New Zealand foraminiferal biostratigraphy (Scott, 1982; Hornibrook and others, 1989). No magnetostratigraphic records extend across this datum in the New Zealand region; thus, accurate determination of its age is not possible at present.

Lack of exposure prevents precise location of the Miocene-Pliocene boundary at Blind

River and Upton Brook. Hilgen and Langereis (1993) show that the Miocene-Pliocene boundary occurs just below the Thvera normal subchron at ca. 5.17 Ma in the Mediterranean. Data from Blind River and Upton Brook support this conclusion. Location of the Thvera subchron in New Zealand sedimentary successions is therefore required to magnetostratigraphically calibrate the position of the Miocene-Pliocene boundary in New Zealand.

Benthonic Foraminifera

***Bolivinita pohana* to *B. pliozea* Transition.** Gradual morphological changes between the benthonic species *B. pliozea* and *B. pohana* occur in a “transition” zone between sites P29/f13 and P29/f29 (230–570 m) at Blind River (Scott, 1978). The *B. pohana* to *B. pliozea* transition is of restricted use for correlation, because it occurs over such a large stratigraphic interval, although the presence of the upper Chron 6 normal subchron in the middle of the interval enables the stratigraphic position to be relatively well constrained (Fig. 13).

Mollusca

First Occurrence Datum of *Sectipecten wollastoni* and *Austrofusus coeruleus*. Kennett (1966b) recorded the first occurrences of the Kapitean index mollusks *S. wollastoni* and *A. coeruleus* at 573 (1,165 m) at Upton Brook. This datum is precisely located stratigraphically and is nearly coincident with the base of Chron 5 (Fig. 11); therefore, the age of the Tongaporutuan-Kapitean boundary is well defined at Upton Brook at ca. 6.38 Ma \pm 0.05 m.y. Reliable magnetostratigraphic records from Kapitean strata have been obtained previously only from sediments of relatively deep-water facies where accurate biostratigraphic location of the Tongaporutuan-Kapitean boundary is not possible. Edwards (1987) concluded from indirect evidence that the boundary lies between 6.2 and 6.7 Ma (see Cande and Kent, 1992). In the New Zealand timescale of Edwards and others (1988), the Tongaporutuan-Kapitean boundary is classified as poorly defined and not useful for external correlation. Magnetostratigraphic results from Upton Brook are the first from strata containing the Kapitean index mollusks and therefore provides the first reliable estimate of the age of the Tongaporutuan-Kapitean boundary.

Last Occurrence Datum of Miocene Mollusks. Extinction of the Kapitean index mollusks and Miocene lineages such as *Cucul-laia* and *Notocorbula* apparently occurred at the end of the time represented by the *G. conomiozea* zone, when 23% of molluscan genera became extinct (Beu, 1990; Beu and Maxwell, 1990). The last occurrence of Miocene mollusks at Upton Brook occurs at the top of the Upton Sandstone at 1,330 m. This datum is estimated to have an age of 5.5 Ma (compare first occurrence datum of *G. sphericomiozea*).

PALEOCEANOGRAPHIC IMPLICATIONS AND CONCLUSIONS

The significant changes in depth of deposition identified by Kennett (1966b) at Blind River and Upton Brook were almost certainly synchronous. Kennett (1967) studied numerous late Miocene and early Pliocene successions in New Zealand, from 46° to 37.5°S, and observed shallowing of sedimentary facies and faunal changes indicative of climatic cooling during late Miocene time. Glacio-eustatic sea level change provides the simplest explanation for synchronous shallowing and climatic cooling over such widely spaced successions (Kennett, 1967). Loutit

and Kennett (1979) and Loutit (1981) inferred that oxygen isotope fluctuations at Blind River were caused by paleoclimatic and glacio-eustatic sea level changes. Edwards (1987) correlated $\delta^{18}\text{O}$ oscillations at Blind River with those recorded at Deep Sea Drilling Project Site 284 and concluded that the oscillations reflect events of regional or global extent, because the localities are some 550 km apart.

Gradual shallowing of the sea is evident in upper Tongaporutuan and Kapitean strata at Blind River and Upton Brook, followed by marked deepening of biofacies and lithofacies at the base of the *G. sphericomiozea* zone (Figs. 2, 11, and 14; Kennett, 1966b). A eustatic sea level rise is inferred to have occurred at the base of the *G. sphericomiozea* zone, because deepening in facies is observed at this datum in all New Zealand successions studied (Beu and Maxwell, 1990). This deepening was followed by shallowing that culminated with shell beds containing *Struthiolaria dolerosa* and *Callusaria obesa* at 1,230 m at Blind River and at 1,400 m at Upton Brook (Figs. 2, 11, and 14; Kennett, 1966b). A marked increase in depth of deposition is evident above these shellbeds at Blind River, as is a smaller increase in depth at Upton Brook (Figs. 2, 11, and 14; Kennett, 1966b).

Oxygen isotope data from the world's oceans indicate that several glacial-interglacial cycles with periods of ~400,000 yr occurred during the late Miocene (Fig. 14; Hodell and others, 1986). High-resolution oxygen isotope records from deep-sea cores indicate two $\delta^{18}\text{O}$ maxima in the Messinian at 5.6 and 5.2 Ma (Keigwin, 1987; compare Cande and Kent, 1992). Although uplift in the Betic-Rif Orogene, due to collision of the African and European plates, is likely to be a major factor in the isolation of the Mediterranean (Weijermars, 1988; Benson and others, 1991; Benson and Rakic-El Bied, 1991), correspondence between the character, timing, and duration of late Miocene events with the Messinian salinity crisis implies that glacio-eustatic sea level events are crucial elements in the salinity crisis (Adams and others, 1977; Hodell and others, 1986). Evidence for the timing of these glacio-eustatic events suggests that the lower Mediterranean evaporite unit was deposited at the same time as the upper part of the *G. conomiozea* zone in the New Zealand region. The marine transgression marked by the first occurrence of *G. sphericomiozea* in the New Zealand region corresponds to the intra-Messinian transgression that separates the Mediterranean evaporite units. Shallow facies sediments in the

upper part of the *G. sphericomiozea* zone correspond to the upper evaporite unit (Fig. 14). By definition, the Miocene-Pliocene boundary is intimately linked to the restoration of marine conditions in the Mediterranean after the Messinian salinity crisis (Cita, 1975; Van Couvering and others, 1976). Shallow marine sedimentation in the late Miocene at Blind River and Upton Brook culminates at the distinctive *Struthiolaria* shell beds that have an inferred age of 5.2 Ma. Deepening of lithofacies and biofacies above these shell beds is inferred to be associated with the basal Pliocene marine transgression (Fig. 14). In Fig. 14, we correlate the New Zealand record to the Mediterranean event stratigraphy of Müller and Hsü (1987), which is tied to the Moroccan stratigraphic control of Hodell and others (1989) and is similar to that proposed by Benson and others (1991) and Benson and Rakic-El Bied (1991). The principal constraints are the correlations of Hodell and others (1989) of the onset of evaporitic conditions in the Mediterranean to the bottom of the upper Chron 5 normal subchron (5.9 Ma) and correlation of a decrease in $\delta^{18}\text{O}$ in the lower Gilbert (5.4 Ma) to the intra-Messinian transgression that marks the base of the upper evaporite sequence (Fig. 14). Precise magnetostratigraphic control on the age of the intra-Messinian transgression is unavailable from Mediterranean sections. The above age estimate of Hodell and others (1989) is based on upward extrapolation of sedimentation rates from the highest polarity boundary recorded in the Moroccan Bou Regreg section. If the $\delta^{18}\text{O}$ decrease recorded at Bou Regreg is due to global ice volume change and resultant sea level rise, then it is likely to be synchronous with the deep-sea $\delta^{18}\text{O}$ record, which indicates interglacial conditions from 5.5 to 5.4 Ma (Hodell and others, 1989; compare Cande and Kent 1992). The basal Zanclean deluge (that is, the Miocene-Pliocene boundary) is placed at 5.17 Ma, following the correlation of Hilgen and Langereis (1993) with the magnetic polarity time scale of Cande and Kent (1992).

The above interpretation is fully consistent with numerous studies that indicate two sea level events in the Mediterranean during the Messinian (for example, Adams and others, 1977; Hsü and others, 1977; Cita, 1982; Cita and McKenzie, 1986). An inferred age of ca. 5.2 Ma (slightly older than the Thvera Subchron) for the Miocene-Pliocene boundary in the New Zealand region is also consistent with the magnetostratigraphic evidence for the age of this datum, as summarized by Hilgen and Langereis (1993).

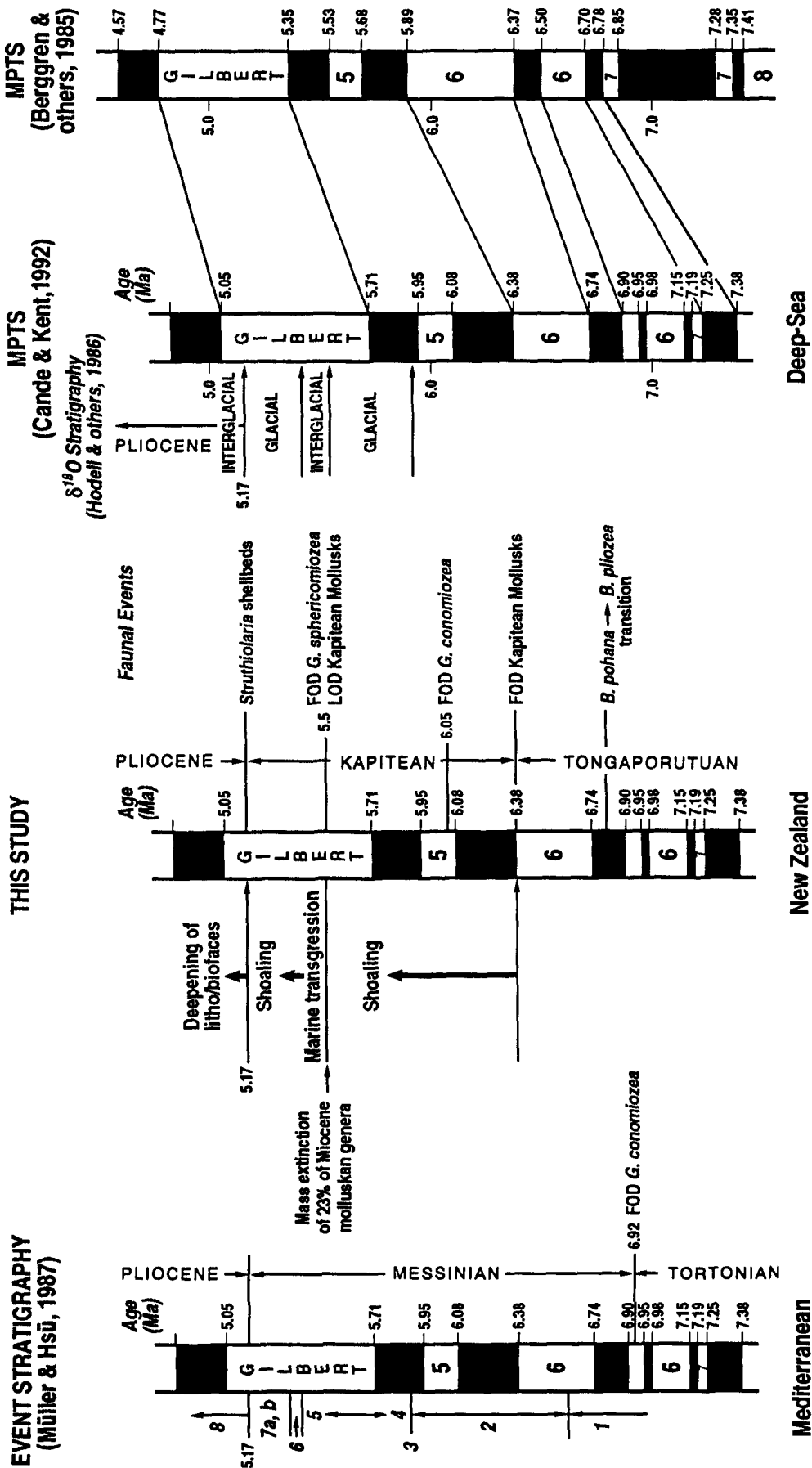


Figure 14. Correlation of terminal Miocene records from New Zealand (this study) with those from the Mediterranean and deep sea. The event stratigraphy of Müller and Hsü (1987) is used for the Mediterranean chronology. 1 = Open marine conditions in the Mediterranean; 2 and 3 = deposition of Tripoli Formation and the Calcare di Base in Sicily; 4 = Main salt deposition; 5 = intra-Messinian desiccation; 6 = intra-Messinian inundation; 7a = upper evaporite; 7b = Lago Mare; and 8 = Zanclean deluge and deposition of Trubi Formation. Position of the Tortonian-Messinian boundary varies from 6.75 to 6.92 Ma (see text), and the position of the Miocene-Pliocene boundary is based on Hilgen and Langerais (1993). The glacial chronology is based on the deep-sea $\delta^{18}\text{O}$ records of Hodell and Kennett (1986), Hodell and others (1986), and Keigwin (1987). The magnetic polarity time scale (MPTS) of Berggren and others (1985) is shown at far right, because most of the papers cited use this time scale, which has been superseded by Cande and Kent (1992). All dates cited are adjusted to the MPTS of Cande and Kent (1992).

Recent revisions of late Miocene and early Pliocene datums enable determination of a high-resolution chronology of terminal Miocene paleoceanographic events. Evidence from around the world suggests a succession of synchronous events of global extent (Fig. 14). Vella (1965) and Loutit and Kennett (1981) proposed that most of the Cenozoic stages in the New Zealand timescale (Fig. 3) represent sedimentary cycles that formed as a result of large, rapid, eustatic sea level events. This is certainly the case in the late Miocene and early Pliocene. Paleontological correlation between New Zealand successions and the international stratotypes is restricted because of the provincial nature of New Zealand faunas. Detailed studies that integrate magneto-, bio-, litho-, and stable isotope stratigraphy should therefore provide a more reliable basis for international correlation.

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