Deglacial paleoceanographic history of the Bay of Plenty, New Zealand

Catherine R. Samson[1]
Institute of Antarctic and Southern ocean Studies, University of Tasmania, Hobart, Tasmania, Australia

Elisabeth L. Sikes
Institute of Marine and Coastal Sciences, Rutgers, The State University of New Jersey, New Brunswick, New Jersey, USA

William R. Howard
Cooperative Research Centre for Antarctic and Southern Ocean Environment, Hobart, Tasmania, Australia

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[1] We present sea surface temperature (SST) records with centennial-scale resolution from the Bay of Plenty, north of New Zealand. Foraminiferal assemblage-based paleo-SST estimates provide a deglacial record of SST since 16.5 ^{14}C ka. Average Holocene SSTs are 15.6°C for winter and 20.3°C for summer, whereas average glacial values were 14.2°C for winter and 19.5°C for summer. Compared to modern time, cooling of SSTs at the Last Glacial Maximum (LGM) was ~0.9°C in winter and ~1.5°C in summer. The shift from glacial to Holocene temperatures began at 14.25 ^{14}C ka, warming by ~2°C until 12.85 ^{14}C ka when temperatures dipped back to glacial values at 11.65 ^{14}C ka. The timing of this return to glacial-like SST correlates well with the Antarctic Cold Reversal (ACR) rather than the Younger Dryas and documents that the influence of the ACR extended into the subtropics of the Southern Hemisphere, at least in this region of the southwest Pacific. By 10.55 ^{14}C ka an SST maximum in summer SSTs of up to 3°C warmer than modern occurred (~24°C), after which SST dropped, remaining at present-day temperatures since 9.3 ^{14}C ka. This early Holocene climatic optimum has been widely noted in the Southern Ocean, and this record indicates that this phenomenon also extended into the subtropics to the north of New Zealand.

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1. Introduction

[2] Understanding the dynamics and forcings of the climate system relies partly on determining critical factors in past climate change. The ice ages which dominated the Pleistocene provide an excellent natural example of climate change. The last deglaciation is of particular interest to climate studies because the timing and sequence of climatic events cannot be explained by orbital forcing alone implying feedbacks are involved in driving rapid climate change [*Lehman and Keigwin*, 1992]. The climate response to the last deglaciation is well documented in the North Atlantic. Pioneering studies there indicated that the deglacial occurred there over several thousand years and consisted of two steps associated with large meltwater events [e.g., *Bard et al.*, 1987; *Jansen and Veum*, 1990; *Ruddiman and McIntyre*, 1981] that were separated by a cooling event, the Younger Dryas. Recent studies on the last deglaciation have shifted from the North Atlantic to regions outside the North Atlantic which provides a more global perspective. Much of the focus has been trying to establish leads and lags between Northern and Southern Hemispheres with the view of better understanding driving forces of climate. Present work suggests a tight interhemispheric coupling of deglacial climate [e.g., *Knutti et al.*, 2004].

[3] The low latitudes appear to play an important role in deglacial climate change [*Guilderson et al.*, 1994; *Howard and Prell*, 1984; *Labeyrie et al.*, 1996; *Lea et al.*, 2000]. Studies have suggested that the tropical temperature response may have led ice volume in the last deglaciation [*Lea et al.*, 2000; *Nürnberg et al.*, 2000] suggesting that the early deglacial warming in the middle-to-high southern latitudes is initiated in the low latitudes [*Howard and Prell*, 1984; *Labeyrie et al.*, 1996; *Lea et al.*, 2000]. Other studies indicate that temperature and moisture transport changes in the tropical Pacific, and in particular the western Pacific warm pool, may drive glacial to interglacial cycles in a manner analogous to the El Niño−Southern Oscillation (ENSO) cycles in the modern ocean [*Koutavas et al.*, 2002; *Stott et al.*, 2002; *Visser et al.*, 2003].

[4] Evidence for a high-latitude Southern Hemisphere lead in deglacial events challenges the view that deglacial climate changes were driven by changes in Northern Hemisphere thermohaline circulation. Antarctic ice core records and marine records from the middle-to-high latitudes of the Southern Ocean suggest that the last deglacial warming occurred 2−4 kyr earlier in the high southern latitudes than the Northern Hemisphere [*Charles et al.*, 1996; *Howard and*

[1]Now at Department of Zoology, University of Tasmania, Hobart, Tasmania, Australia.

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Prell, 1984, 1992; *Labeyrie et al.*, 1996; *Mashiotta et al.*, 1999; *Pichon et al.*, 1992; *Sowers and Bender*, 1995]. Overall, different temperature proxies globally are contradictory. For example, tropical SST changes based on Sr/Ca in Barbados corals appear synchronous with variations in planktonic δ^{18}O from the South Atlantic Ocean and Antarctic Vostok deuterium records [*Charles et al.*, 1996] implying synchroneity between the low latitudes and high southern latitudes. In contrast, an alkenone SST record from the tropical Indian Ocean suggests climatic changes in the low southern latitudes were synchronous with those in the Northern Hemisphere [*Bard et al.*, 1997]. In the Equatorial Atlantic different proxies also show different timing relative to ice volume [*Nürnberg et al.*, 2000; *Sikes and Keigwin*, 1994].

[5] The Younger Dryas cold event was originally thought to occur only in the North Atlantic region but several glacial, vegetational, and isotopic records suggested worldwide evidence for Younger Dryas correlatives. A global Younger Dryas event challenges the view that reduced NADW production is the only driving mechanism and although evidence exists for a global Younger Dryas event beyond the North Atlantic their correlation and interpretation are controversial [*Peteet*, 1995; *Rind et al.*, 1986]. Palynological records indicate cooling in both North and South America [*Peteet*, 1995]. In New Zealand and South America, although glacial readvances suggest a Younger Dryas correlative [*Denton and Hendy*, 1994; *Heusser and Rabassa*, 1987], the pollen records are equivocal [*Markgraf*, 1993; *McGlone*, 1995] and dating and interpretation of the glacial advance in New Zealand remains controversial [*Fitzsimons*, 1997; *McGlone*, 1995].

[6] Low- to middle-latitude records from the around the northern and eastern equatorial Pacific, Sulu Sea, and Gulf of Mexico have been interpreted as showing a Younger Dryas correlative with a distinct δ^{18}O enrichment of 0.4–0.8‰ between $\sim$11 and 10 ka [*Flower and Kennett*, 1990; *Hendy and Kennett*, 1999; *Kallel et al.*, 1988; *Keigwin and Jones*, 1990; *Kennett and Ingram*, 1995; *Kudrass et al.*, 1991; *Linsley and Thunell*, 1990; *Stott et al.*, 2002]. This δ^{18}O enrichment, combined with changes in planktonic foraminiferal assemblages, is evidence for a Younger Dryas event in these locations, but may indicate a lowered salinity or atmospheric transport of the meltwater signal rather than cooling [*Anderson and Thunell*, 1993; *Keigwin and Gorbarenko*, 1992; *Thunell and Miao*, 1996]. In contrast, tropical records from the western equatorial and Southern Hemisphere tropical Pacific show no Younger Dryas correlative [*Lea et al.*, 2000; *Visser et al.*, 2003]. Previous paleoceanographic work in the Bay of Plenty indicates a last glacial cooling of $\sim$2°C, but these results are of too low resolution to resolve rapid climate changes such as the Younger Dryas [*Wright et al.*, 1995]. To date no one has produced a century-scale record from the Pacific Southern Hemisphere subtropics.

[7] Most Antarctic ice cores exhibit a late deglacial cooling, marked by a decrease in deuterium and oxygen isotope values [*Jouzel et al.*, 1987, 1992, 1995]. This oscillation, known as the Antarctic Cold Reversal (ACR) occurs at 14–12.5 calendar (cal) kyr B.P. [*Jouzel et al.*, 2001] (12.4–11.1 ^{14}C ka) and precedes the Younger Dryas by approximately 1 kyr [*Blunier et al.*, 1997; *Sowers and Bender*, 1995]. In the Southern Ocean, this cold oscillation seen in ice cores appears synchronous with a late deglacial enrichment in oxygen isotope records from the Indian sector [*Labracherie et al.*, 1989] and is seen in high-resolution records in subantarctic waters east of the South Island of New Zealand [*Pahnke et al.*, 2003]. This suggests that the presence of the ACR may be typical of the high latitudes of the Southern Hemisphere [*Stenni et al.*, 2001]. This climatic link between the ACR and the Northern Hemisphere Younger Dryas may be due to a bipolar seesaw effect stemming from freshwater discharge [*Knutti et al.*, 2004; *Stocker and Johnsen*, 2003].

[8] The Bay of Plenty, in the subtropical southwest Pacific, to the north of New Zealand, provides an excellent location for studying low-latitude Southern Hemisphere response to deglacial forcing. The Bay of Plenty sits proximal to the western Pacific warm pool and shows a modern day climatic response to ENSO events. Because of its close proximity to the New Zealand landmass, the sediments contain pollen as well as foraminifera thus enabling comparison of the oceanic and atmospheric/terrestrial response to deglaciation. In addition, the Bay of Plenty lies proximal to the region of active volcanism associated with the Taupo Volcanic Zone (TVZ) (Figure 1). The Taupo Volcanic Zone has frequently and regularly erupted throughout the late Quaternary placing numerous rhyolitic tephras in the marine sediments in the area [*Pillans and Wright*, 1992; *Wright et al.*, 1990]. Late Quaternary sedimentation in the bay is primarily hemipelagic, interspersed with horizons of tephras which are often shower bedded indicating that they are deposited directly from the ash cloud and not transported via marine processes such as turbidity currents [*Kohn and Glasby*, 1978]. Tephras in the sediments in the Bay of Plenty have been correlated to their radiocarbon-dated counterparts on land via glass chemistry and heavy mineral assemblages [*Kohn and Glasby*, 1978] and provide stratigraphic tie points which improve the chronology in marine cores [*Sikes et al.*, 2000].

[9] Surface water in the Bay of Plenty is subtropical, with modern mean summer and winter SSTs of 21.0°C and 15.1°C, respectively [*Bottomley et al.*, 1990]. Regional subtropical surface waters originate in the central Pacific Ocean, then flow westward in the South Equatorial Current before heading southward along the east coast of Australia (East Australian Current). General east to west movement of waters occurs across the Tasman Sea and around the northern tip of New Zealand [*Tomczak and Godfrey*, 1994]. At the northern tip of New Zealand, the East Auckland current coalesces and flows southeastward around northern New Zealand and through the Bay of Plenty over the H214 core site.

[10] To better constrain the relative timing of deglacial climate change in the mid latitudes of the Southern Hemisphere, this study presents a centennial-scale deglacial record of sea surface temperature (SST) changes for the Bay of Plenty using planktonic foraminifera from core H214 (36°55.5′S, 177°26.5′E, 2045 m water depth). We compare the SST record with the pollen record from another marine core [*Wright et al.*, 1995] to establish the relative

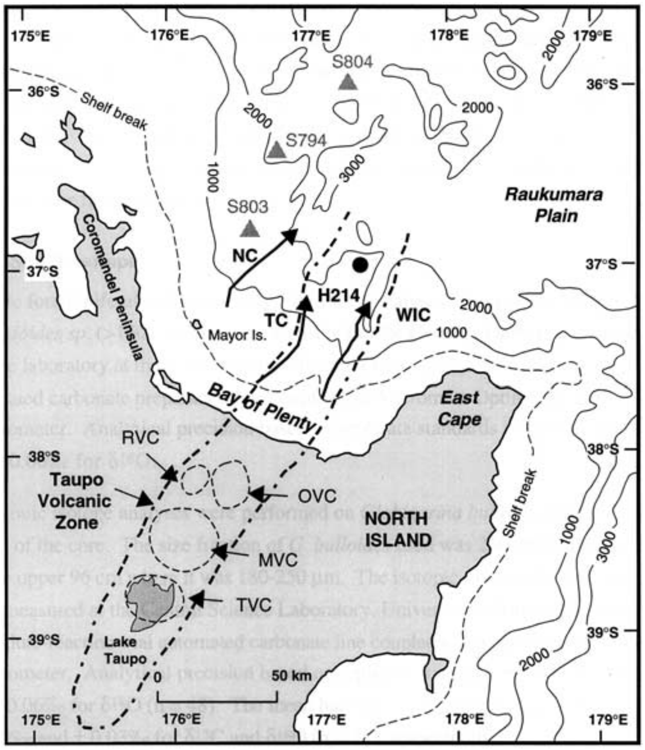

Figure 1. Bathymetric map of the Bay of Plenty showing the location of core H214 (solid circle) and other cores mentioned in the text (grey triangles). Isobaths are in meters. Three submarine canyons, Ngatoro Canyon (NC), Tauranga Canyon (TC), and White Island Canyon (WIC), extend from the shelf break across the continental slope and rise. The Taupo Volcanic Zone with its constituent volcanic centers, the Maroa Volcanic Center (MVC), the Okataina Volcanic Center (OVC), the Rotorua Volcanic Center (RVC), and the Taupo Volcanic Center (TVC), is the predominant source of rhyolitic tephra found in the Bay of Plenty sediments. After *Pillans and Wright* [1992].

timing of deglacial response in the oceanic and terrestrial (atmospheric) systems.

2. Methods

[11] Core H214 was sampled at ~5 cm intervals between 0 and 102 cm and 211 and 262 cm and every centimeter between 102 and 211 cm. Samples were oven dried, disaggregated and wet sieved to 63 μm. Benthic foraminiferal carbon and oxygen isotope analyses were performed on *Cibicidoides sp*. (>150 μm) and were conducted by H. Spero at the Geology Department, University of California, Davis, on a common-acid-bath automated carbonate preparation line coupled to a Micromass Optima mass spectrometer. Analytical precision based on replicate standards is ± 0.03‰ for δ^{13}C and ± 0.06‰ for δ^{18}O. Planktonic isotope analyses were performed on *Globigerina bulloides* (250–300 μm below 96 cm and above, from the 180–250 μm fraction). Planktonic analyses were conducted at the Central Science Laboratory, University of Tasmania, on an individual-reaction-vial automated carbonate line coupled to a Micromass Optima mass spectrometer. Analytical precision based on replicate standards is ± 0.03‰ for δ^{13}C and ± 0.06‰ for δ^{18}O (n = 48). The mean half range for replicate samples is ± 0.09‰ and ± 0.03‰ for δ^{13}C and δ^{18}O (n = 11), respectively. Because *G. bulloides* abundances were low in the upper portion of the core, isotopes were also run on *Globigerinoides ruber* in the upper 164 cm of the core (the 250–300 μm size fraction below 27 cm and above, 180–250 μm). Samples in the upper 107 cm of the core were run at the University of Tasmania (~30 tests) and samples below 107 cm were run at the University of California, Davis (~5–10 tests). A number of samples were run at both locations and the isotopic results are identical, within the error. Prior to analysis, foraminifera were sonicated in methanol, crushed in methanol and roasted under vacuum for one hour at 370°C to remove any organic matter. Planktonic and benthic results are reported as per mil (‰) deviations from the Peedee belemnite (PDB) standard using Carerra marble as a laboratory standard [*Samson*, 1998].

[12] Samples for foraminiferal assemblage temperature estimation were sieved to ≥150 μm and successively split, in a microsplitter, until 300–600 whole planktonic foraminifera were obtained for faunal analyses. Planktonic foraminiferal species were classified following the taxonomy used by *Prell* [1985], based on the work of *Bé* [1977], *Kipp* [1976], and *Parker* [1962]. For faunal analyses, the 29 species and morphotypes recognized by *Kipp* [1976] and used by *CLIMAP Project Members* [1976, 1981] were counted with the exception that *Neogloboquadrina pachyderma* (dextral)–*N. dutertrei* intergrade is no longer counted as a separate taxon [*Prell et al.*, 1999]. Of the twenty species identified in this core, one species, *Pulleniatina obliquiloculata*, made up <1% in all samples and was therefore excluded from the species list for modern analog SST estimates. SST estimates were then obtained using the modern analog technique [*Anderson et al.*, 1989; *Howard and Prell*, 1992; *Overpeck et al.*, 1985; *Prell*, 1985].

[13] The modern analog technique matches a down-core assemblage with modern core top samples that have similar faunas [*Prell*, 1985]. The 10 best fit core tops are chosen using squared chord distance, and the weighted (using a corresponding squared chord similarity) average of their associated temperatures is the temperature estimate. In this study the modern SST climatology is the "GOSTA" data set [*Bottomley et al.*, 1990]. The method permits a sample by sample estimate of the fit between down-core samples and modern core tops. Ancient samples with close modern analogs have low dissimilarity coefficients. This comprises one measure of the reliability of the estimate. Dissimilarity coefficients (squared chord distance) of zero are considered a perfect match, whereas a value of 2 is considered completely dissimilar. The dissimilarity coefficients are generally excellent: Most values are below 0.2; all values are less than 0.3 (with the exception of the core top sample) which are considered acceptable matches (Table 2). The modern core top foraminiferal counts database used is the

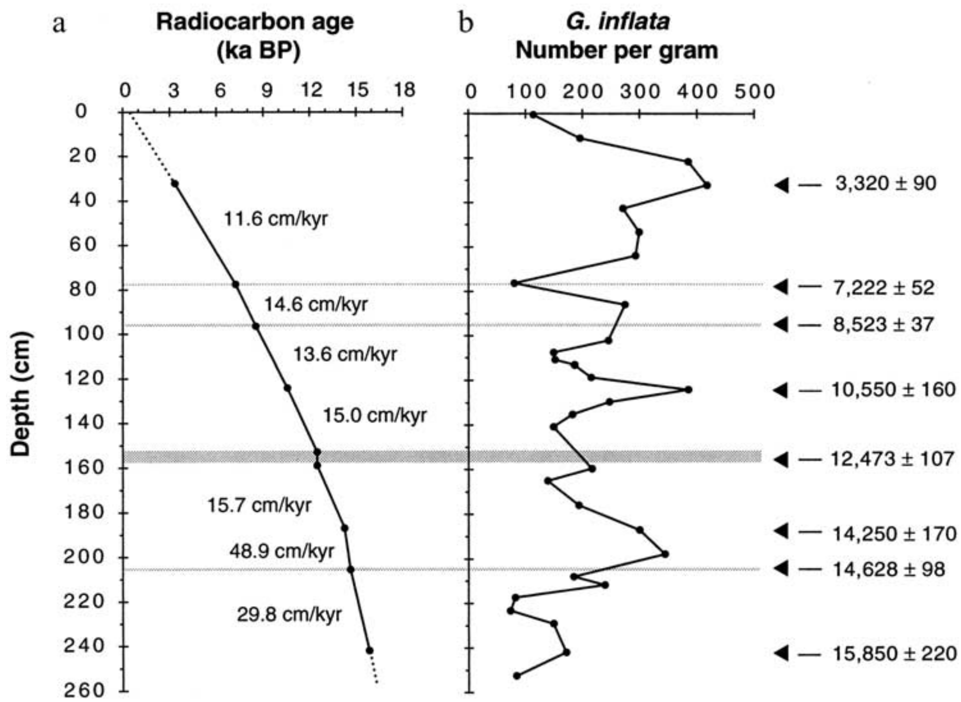

Figure 2. Construction of the marine ^{14}C timescale for core H214. (a) Age versus depth plot with the sedimentation rates inferred from age/depth relations. The dotted lines mark sedimentation rates extrapolated for generating a continuous chronology. (b) Accelerated mass spectrometry (AMS) ^{14}C levels relative to *Globorotalia inflata* abundances (number per gram). The AMS ^{14}C ages of the tephras, marked by grey bands, are an average of dates from directly above and below the tephras. AMS ^{14}C dates not associated with the tephras were taken at local maxima in *G. inflata* abundance. The rhyolitic tephras were identified by *Kohn and Glasby* [1978], whose results, combined with marine ^{14}C dating by *Sikes et al.* [2000], identify the ashes as the Mamaku 7.25 ka (77–78 cm), Rotoma 8.53 ka (96–97 cm), Waiohau 11.58 ka (153–159 cm), and Rerewhakaaitu 14.7 ka (205–206 cm) tephras [*Froggatt and Lowe*, 1990].

same as that used by CLIMAP except for the exclusion of *Globorotalia crassiformis* and *Neogloboquadrina pachyderma* (dextral)–*N. dutertrei* intergrade in the assemblage counts [*Prell et al.*, 1999]. Compared with CLIMAP estimates, analog estimates yield correlations equal or better to observed SST and have lower standard errors [*Prell*, 1985]. *Prell* [1985] showed that modern analog estimates reproduce the observed SST better than estimates based on a foraminiferal-based transfer function equation for the Pacific. The modern analog technique works directly with species percentage data and does not require any factor analysis which may smooth or generalize the data [*Prell*, 1985]. This technique provides a standard deviation for each SST estimate based on the range of values in the subset, and the analog samples provide geographic information which can be used to interpret the past environment of the subject sample [*Prell*, 1985].

3. Results

3.1. Stratigraphy and Chronology

[14] Core H214, retrieved from 2045 m water depth sits removed from the submarine canyon systems in the Bay of Plenty (Figure 1). The core appears to have continuous sedimentation unaffected by local turbidity currents [*Pillans and Wright*, 1992; *Wright et al.*, 1990]. The sediment consists of fine-grained hemipelagic ooze interspersed with four pale grey rhyolitic tephras and two black andesitic tephras [*Kohn and Glasby*, 1978]. The distribution of tephras in Bay of Plenty cores is well characterized and we follow the accepted regional practice of making the final identification for ashes in a core based on a combination of relative position in the core, identification of the ferromagnesian assemblage, the chemical variations within the titanomagnetites [*Kohn and Glasby*, 1978], and their radiocarbon ages [*Pillans and Wright*, 1992; *Sikes et al.*, 2000]. The rhyolitic tephras and ^{14}C ages associated with them were used as stratigraphic tie points within the core (Figure 2).

[15] The final chronology adopted for core H214 is based on a combination of oxygen isotope stratigraphy (Figure 3), additional ^{14}C ages (Figure 2) and the position of the tephras. Accelerator mass spectrometry (AMS) ^{14}C dates not associated with tephras were performed on the planktonic foraminifer *Globorotalia inflata* at peaks in their abundance (Table 1 and Figure 2). Sedimentation rates were linearly interpreted between AMS ^{14}C dates. Although the H214 record does not extend to the Last Glacial Maximum (LGM) at ~20–18 ka, the magnitude of the planktonic and benthic oxygen isotopic shift is equivalent to the full glacial-interglacial signal observed in other cores from the

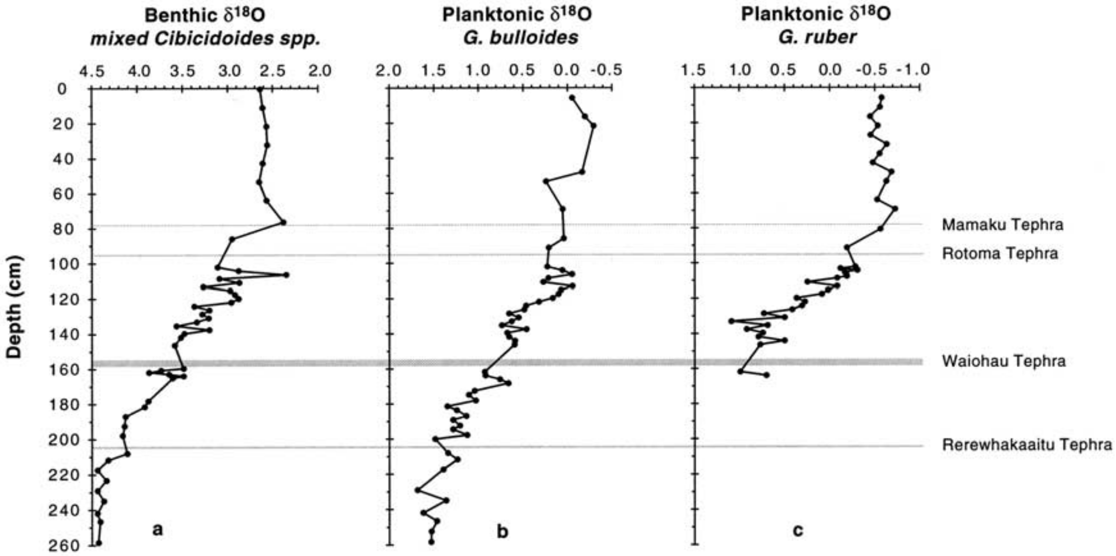

Figure 3. Oxygen isotopes from core H214 versus depth. (a) Benthic oxygen isotopes. Analyses were run on mixed species of the genus *Cibicidoides.* (b) Oxygen isotopes for the planktonic foraminifer *Globigerina bulloides.* (c) Oxygen isotope results for the planktonic species *Globigerinoides ruber.* Grey bands represent the levels at which the respective tephras were located: 77–78 cm, Mamaku; 96–97 cm, Rotoma; 153–159 cm, Waiohau; and 205–206 cm, Rerewhakaaitu [*Froggatt and Lowe*, 1990].

region [*Newnham et al.*, 2003; *Wright et al.*, 1995], indicating that the entire deglacial transition is recorded in H214. A constant reservoir correction of −400 years was applied to all marine AMS ^{14}C ages to account for the apparent ^{14}C age of low-latitude surface waters [*Bard et al.*, 1988]. All ages reported here for core H214 are in ^{14}C years before present (^{14}C ka), unless otherwise noted.

[16] Sedimentation rates were highest during the late glaciation and early deglaciation (~30–50 cm/kyr) (Figure 2). After ~14.25 ka sedimentation rates decreased to ~11.5–15.5 cm/kyr and remained approximately constant for the remainder of the deglaciation and during the Holocene. Higher sedimentation rates during glacial intervals are characteristic of offshore sedimentation in northern

Table 1. List of Accelerator Mass Spectrometry (AMS) ^{14}C Dates for Core H214

Core Depth, cm	Ash Layer	Foraminiferal Species	Size Fraction, μm	^{14}C Age, yr B.P.	Corrected Age,[a] yr B.P.	1 σ Error	AMS Lab Number
32.2	-	G. *inflata*	>150	3,720	3,320	90	OZD 261
75–77		G. *inflata*	>150	7,637	7,237	87	NZA 6654
76–77		G. *inflata*	>150	7,540	7,140	40	LL 39597
77–78	Mamaku	-	-	-	-	-	-
78–79		G. *inflata*	>250	7,610	7,210	40	LL 39598
78–79		G. *inflata*	>250	7,700	7,300	40	LL 39599
94–96		G. *inflata*	>150	8,800	8,400	30	LL 39600
96–97	Rotoma	-	-	-	-	-	-
97–98		G. *inflata*	>250	9,130	8,730	40	LL 39601
97–98		G. *inflata*	>250	8,960	8,560	40	LL 39602
124.2	-	G. *inflata*	>250	10,950	10,450	160	OZD 262
151–153		G. *inflata*	>150	12,820	12,420	110	NZA 6655
153–159	Waiohau	-	-	-	-	-	-
159–160		G. *inflata*	>250	12,910	12,510	140	NZA 6662
160–161		G. *inflata*	>250	12,940	12,540	70	LL 40465
186.9	-	G. *inflata*	>250	14,650	14,250	170	OZD 263
202–203		G. *inflata*	>250	14,980	14,580	70	LL 40466
203–205		G. *inflata*	>150	14,980	14,580	120	NZA 6663
205–206	Rerewhakaaitu	-	-	-	-	-	-
206–208		G. *inflata*	>150	15,350	14,950	160	NZA 6664
208–209		G. *inflata*	>250	14,970	14,570	70	LL 40467
208–209		G. *inflata*	>250	14,910	14,510	70	LL 40468
241.9	-	G. *inflata*	>250	16,250	15,850	220	OZD 264

[a]A correction of 400 years was applied to account for the ^{14}C surface reservoir age [*Bard et al.*,1988]. All dates were derived from monospecific samples of the planktonic foraminifer *Globorotalia injlata.* Samples with lab numbers beginning with NZA were run at the Rafter Radiocarbon Laboratory, New Zealand. Those beginning with OZD were run at the Australian Nuclear Sciences and Technology Organisation (ANSTO), and those beginning with LL were run at the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory, California.

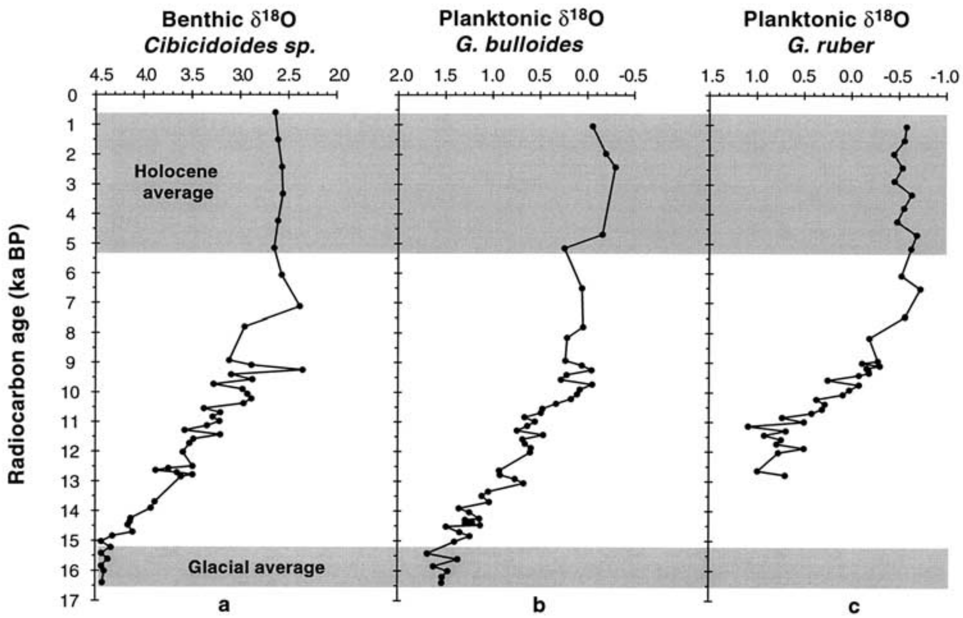

Figure 4. Oxygen isotopes from core H214 versus age. (a) Benthic oxygen isotopes. Analyses were run on mixed species of the genus *Cibicidoides.* (b) Oxygen isotopes for the planktonic foraminifer *Globigerina bulloides.* (c) Oxygen isotope results for the planktonic species *Globigerinoides ruber.* The top grey band marks the interval used as average Holocene values, while the bottom grey band marks the interval used as average glacial values.

and eastern New Zealand because of increased supply of glaciofluvial sediment to the deep sea as a result of lowered sea level and increased erosion in alpine regions [e.g., *Sikes et al.*, 2002; *Wright et al.*, 1995].

3.2. Stable Isotopes

[17] An estimate of the overall magnitude of the glacial-Holocene planktonic isotopic shift was obtained by averaging glacial and Holocene isotopic values. Depths for averaging were selected on the basis of benthic $\delta^{18}O$ values. The glacial and Holocene averages are the mean of the isotopic values between ~16.4–15.2 ka and ~5.1–0.6 ka, respectively (Figure 4). All numerical data (isotopic, assemblage, etc.) are archived and available online from the World Data Center for Paleoclimatology (available at http://www.ngdc.noaa.gov/paleo).

[18] Average glacial benthic $\delta^{18}O$ values were 4.4‰ and average Holocene values are 2.61‰ with the magnitude of the glacial-interglacial shift being 1.79‰. The deglacial transition, marked by decreasing $\delta^{18}O$ values, occurs between ~15.0 ka and 7.1 ka (~217–76 cm). Average glacial planktonic $\delta^{18}O$ values for *G. bulloides* were 1.54‰ and average Holocene values are −0.09‰, giving a glacial to Holocene shift of ~1.63‰. For planktonics, the deglacial transition begins at the same time as the benthic shift but finishes later at ~7.5 ka. Average Holocene planktonic $\delta^{18}O$ values for *G. ruber* are −0.55‰. *G. ruber* abundances are low below ~12.8 ka, whereas *G. bulloides* numbers are low above ~8 ka. To obtain a reliable, high-resolution planktonic isotopic record for the length of the core, both species were run [*Samson*, 1998] (Figure 4). The planktonic isotopic signals are very similar but differ slightly both in the deglaciation and the Holocene; this may be due in part to different habitat and seasonalities in growth between the species.

3.3. SST Estimates

[19] Holocene and glacial averages for SST were calculated from the same levels chosen for isotopic averages. Average Holocene values for H214 are 15.6°C for winter and 20.3°C for summer. These accurately reflect, within the error of the estimates, modern winter and summer SSTs at the site which are 15.1°C and 21.0°C, respectively [*Bottomley et al.*, 1990]. Average glacial values were 14.2°C for winter and 19.5°C for summer suggesting SSTs at the LGM were ~0.9°C colder in the winter and ~1.5°C colder in the summer than modern time (Figure 5a).

[20] SSTs remained at cooler "glacial" temperatures until ~14.25 ka (Figure 5a). Between ~14.25 ka and ~13.5 ka temperatures warmed by ~2°C, and SSTs reached Holocene values between ~13.5 and ~12.85 ka. At ~12.85 ka temperatures began to decrease and by ~11.65 ka temperatures had returned to glacial values. Temperatures steadily increased between ~11.65 and 10.55 ka. A maximum is clearly evident in summer SSTs between ~10.55 and 9.3 ka, with summer SSTs up to 3°C warmer than today. After ~9.3 ka SSTs were relatively stable at ~20.5°C for summer and ~15.7°C for winter. There may have been a second smaller temperature maximum ~6.0 ka but it is a single point peak and further data are required to substantiate this result.

[21] In all but two samples (i.e., 107.5 cm and 113 cm) the mean dissimilarity coefficients are less than 0.2, indicating that down-core faunal assemblages have good modern analogs (Table 2 and Figure 5b). As the dissimilarity values for the two samples are only marginally greater than 0.2 and the temperature estimates are compatible with neighboring

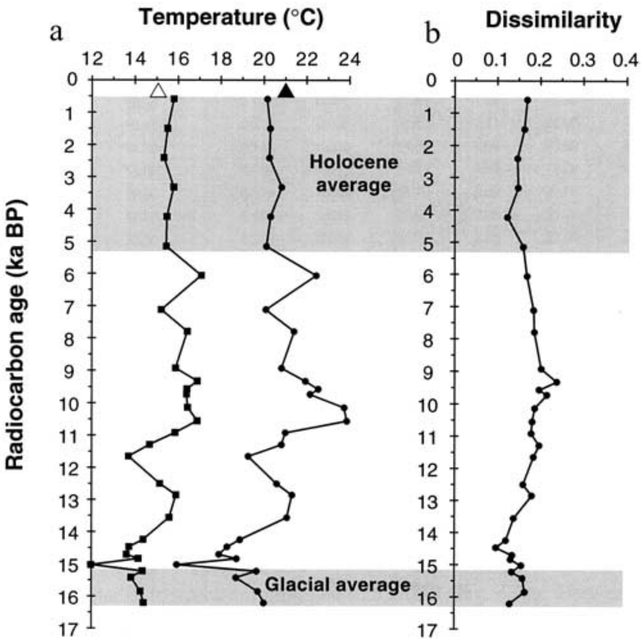

Figure 5. (a) Cold season (squares) and warm season (circles) sea surface temperature estimates versus age for core H214. Modern cold season and warm season SST [*Bottomley et al.*, 1990] are marked by an open triangle and closed triangle, respectively. (b) Dissimilarity coefficients for the assemblages. These provide a measure of the reliability of the temperature estimate and for this core are considered acceptable. The top grey band marks the interval used as average Holocene values, while the bottom grey band marks the interval used as average glacial values. Average glacial SSTs were about 1°–1.5°C colder than modern SSTs. As the H214 record only extends back to 16.4 ka, the glacial-Holocene temperature difference obtained from biotic estimates in core H214 must be considered a minimum.

samples, it seems likely that the temperature estimates for the two samples are reliable. The standard deviations associated with the SST estimates are about 1.5°C for the Holocene and about 2°–3°C for deglacial SST estimates (Table 2).

4. Discussion

4.1. Amplitude and Timing of Deglacial Warming

[22] Assemblage-based SSTs in core H214 indicate temperatures during the last glaciation were at least 1.5°C colder in summer and 0.9°C colder in winter than today. Core H214 only extends back to 16.4 ka so, although it appears to record a full glacial-interglacial oxygen isotopic transition, it may not record the coldest glacial temperatures. Pollen records from marine cores in the Bay of Plenty indicate that the coldest temperatures occurred early in marine isotope stage 2, immediately before the deposition of the Kawakawa tephra at ~22.59 ka [*Wright et al.*, 1995]. Therefore the glacial-Holocene temperature differences obtained from biotic estimates in core H214 must be considered a minimum.

[23] SSTs in H214 began their initial warming from glacial levels at 15.5 ka with a broad interval of warming occurring directly after the appearance of the Rerewhakaaitu ash in the core at ~14.5 ka and continuing until ~13.5 ka (Figure 6b). The timing of the initiation of the broad warming in H214, occurring directly after the Rerewhakaaitu, is widely observed in the New Zealand region where this ash is considered an important stratigraphic marker for the deglaciation [*Newnham et al.*, 2003].

[24] Maximum temperatures in H214 occurred between ~10.55 and 9.3 ka, centered on ~10.5 ka. This 1000 year warming, with temperatures as much as 3°C warmer than the Holocene, is roughly synchronous with the early Holocene temperature maximum or "climatic optimum" that is widespread in records from the Indian and Australian–New Zealand region of the Southern Ocean and Antarctic ice core records [e.g., *Ciais et al.*, 1992, 1994; *Hutson*, 1980; *Ikehara et al.*, 1997; *Labeyrie et al.*, 1996; *Labracherie et al.*, 1989; *Morley*, 1989; *Sikes et al.*, 2002; *Stenni et al.*, 2001; *Weaver et al.*, 1998; *Wells and Okada*, 1996, 1997]. This is the first evidence of this early Holocene temperature maximum in subtropical waters in the region [*Nelson et al.*, 2000; *Sikes et al.*, 2002] and suggests a stronger link to sub Antarctic/Antarctic climate in the New Zealand area than previously suggested [*Stenni et al.*, 2001].

[25] The magnitude of isotopically derived LGM-Holocene temperature differences is strongly dependent on the choice of the estimate for the glacial-interglacial effect of global ice volume on the $\delta^{18}O$ signal. The best estimate of the global ice volume effect is 1.0‰ [*Schrag et al.*, 1996]. The planktonic $\delta^{18}O$ glacial-interglacial amplitude in core H214 is 1.63‰. Using the *Schrag et al.* [1996] estimate, the glacial-interglacial planktonic $\delta^{18}O$ shift in H214 suggests LGM temperatures were ~2.8°C colder than today. Using planktonic oxygen isotopes in Bay of Plenty cores, *Wright et al.* [1995] reported that LGM SSTs were 2°–3°C colder than today in three cores from the Bay of Plenty (Figure 1). However, *Wright et al.* [1995] employed a global ice volume effect range of 1.33‰ [*Fairbanks*, 1989; *Fairbanks and Matthews*, 1978] because the *Schrag et al.* [1996] estimate postdates that study. Using the *Fairbanks* [1989] estimate for ice volume effect, the glacial-interglacial LGM temperature change would be ~1.4°C than today, a slightly lower value than *Wright et al.* [1995], but in agreement with the assemblage-based estimates in H214. Significantly, SST estimates based on Mg/Ca from the western tropical Pacific indicate a glacial-interglacial temperature range of 3°–4°C [*Visser et al.*, 2003]. Results from the Indonesian region suggest that initial temperature warming had already occurred by 16 ^{14}C ka and that SST warmed 2000–3000 years before the decrease in ice volume was reflected in the $\delta^{18}O$ record. This supports our interpretation that in H214, the coldest temperatures of the LGM were not obtained and suggests the greater temperature range seen in the isotopic record of H214 is a reflection of the persistence of an ice volume signal in the early deglaciation [*Visser et al.*, 2003].

4.2. Marine versus Terrestrial Record of Deglaciation

[26] The pollen record from core S803 in the Bay of Plenty (Figure 1) provides a marine-derived terrestrial record of deglacial climate change from northern New

Table 2. Planktonic Foraminiferal Warm Season and Cold Season SST Estimates Derived Using the Modern Analog Technique for Core H214

Core Depth, cm	Radiocarbon Age, ka	Dissimilarity[a]		SST Cold Season		SST Warm Season	
		Average	Standard Deviation		Standard Deviation	Average	Standard Deviation
0.5	0.59	0.169	0.037	15.80	1.63	20.14	1.68
11.1	1.5	0.163	0.025	15.52	1.50	20.3	1.55
21.7	2.42	0.147	0.021	15.34	1.45	20.26	1.55
32.2	3.32	0.148	0.027	15.79	1.48	20.83	1.74
42.8	4.23	0.124	0.032	15.48	1.60	20.31	1.84
53.4	5.15	0.161	0.017	15.45	1.62	20.11	1.91
64	6.06	0.168	0.022	17.08	1.18	22.44	1.54
76.3	7.12	0.184	0.013	15.22	1.26	20.09	1.24
85.9	7.79	0.186	0.03	16.43	1.66	21.41	1.79
102	8.92	0.202	0.025	15.89	1.53	20.82	1.34
107.5	9.33	0.238	0.026	16.90	1.33	21.94	1.16
110.9	9.58	0.197	0.013	16.42	1.57	22.54	1.82
113.1	9.74	0.215	0.012	16.40	1.57	22.16	1.86
118.6	10.14	0.187	0.021	16.45	2.04	23.74	2.74
124.2	10.55	0.181	0.025	16.90	1.56	23.87	3.31
129.7	10.92	0.179	0.008	15.87	2.41	21.00	3.00
135.3	11.29	0.198	0.023	14.69	2.06	20.82	2.81
140.8	11.66	0.184	0.015	13.71	1.26	19.28	2.27
159.5	12.5	0.16	0.012	15.16	2.60	20.60	3.46
165	12.85	0.18	0.015	15.92	2.98	21.32	3.28
176	13.55	0.139	0.007	15.60	2.48	21.08	3.44
186.9	14.25	0.12	0.018	14.40	1.94	18.90	2.11
197.9	14.47	0.097	0.011	13.74	1.46	18.31	1.68
207.9	14.71	0.134	0.009	13.63	2.19	17.92	2.95
211.6	14.83	0.131	0.018	14.17	2.00	18.75	2.18
217.3	15.03	0.156	0.017	11.99	4.01	15.96	4.02
223.2	15.22	0.134	0.024	14.36	3.22	19.68	2.69
229	15.42	0.159	0.014	13.84	2.23	18.71	2.35
241.9	15.85	0.165	0.02	14.28	3.41	19.73	3.02
252.5	16.21	0.129	0.018	14.42	3.20	20.02	2.46

[a]Dissimilarity values are less than 0.2 indicating that the H214 assemblages have good modern analogs.

Zealand [*Wright et al.*, 1995]. Although foraminiferal SST estimates in H214 and pollen records from core S803 imply a similar deglacial response, the timing differs between the two proxies (Figure 6). The pollen record suggests that climate conditions warmed and moistened between ~16.5 and 14.6 ka and a return toward glacial conditions occurs in the pollen record from 14.6 to 12 ka when increases in conifers, hardwoods and tree ferns suggest climatic conditions were becoming warmer and wetter. The timing and climatic shift correlates well with pollen records from the North Island (summarized by *Newnham et al.* [2003]). In contrast, the SST record suggests climatic conditions did not begin to ameliorate until ~14.5 ka (unless the one point spike at 15 ka in the SST record is taken into account) and SSTs did not begin to cool until ~12.85 ka. The second phase of warming appears synchronous in the marine and pollen records occurring at ~12–10 ka. Early in the deglaciation climatic changes appear to occur about 2000 years earlier in the pollen record than the SST record whereas after ~12 ka the response appears to be synchronous in the two records. SST and pollen records both suggest slightly warmer conditions in the early Holocene than during the late Holocene.

[27] Could the differences in timing between the SST and pollen records be an artifact of the chronologies applied to the records? The H214 chronology is based on a reservoir-corrected marine ^{14}C timescale, assuming a constant surface reservoir age of 400 years. The chronology for S803 is based on a combination of marine ^{14}C ages, corrected by −400 years to account for the reservoir effect and in the early deglaciation, during the interval in which the timing in H214 and S803 disagrees, terrestrial ^{14}C ages of the tephras. Marine and terrestrial ^{14}C derived chronologies will produce a common timescale only if the assumption of a constant reservoir age holds true. The differences between the H214 marine and terrestrially based ^{14}C timescales may be partly due to temporal changes in surface reservoir ages during the deglaciation [*Sikes et al.*, 2000].

[28] As correlatives of the tephras in core H214 have been ^{14}C dated on land [*Froggatt and Lowe*, 1990] a terrestrially based ^{14}C chronology was developed for H214 (Figure 6b) to test whether the apparent differences in timing between H214 SST and S803 pollen records are an artifact of using a marine versus a terrestrially based ^{14}C chronology. When compared on a common terrestrially based ^{14}C timescale the S803 pollen and H214 SST records still remain offset by ~2 kyr early in the deglaciation with SST lagging terrestrial warming (Figure 6c). Ideally, to compare the timing of changes in pollen and SSTs the records would be from the same core to ensure they were on a common chronology and a common stratigraphy. However, the offset in the timing of deglacial climatic change in the H214 SST record and the S803 pollen record is too large to be an artifact of core chronologies. Taken at face value, the earlier response in the pollen record compared to the SST record suggests

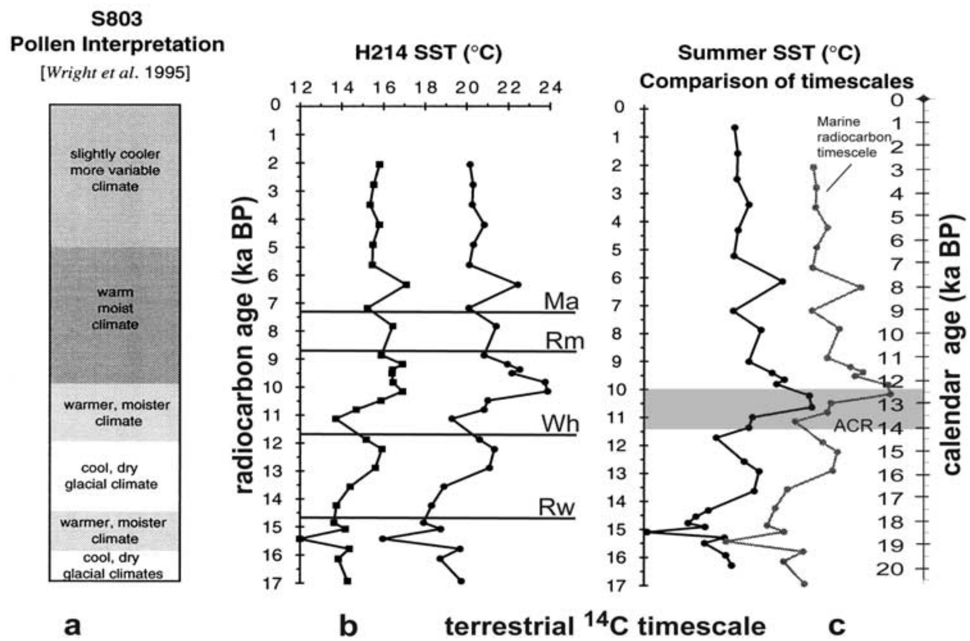

Figure 6. Sea surface temperature and pollen records of deglaciation in the subtropical Pacific. (a) Deglacial climate change derived from pollen from core S803 [*Wright et al.*, 1995]. (b) H214 summer (circles) and winter (squares) SST record presented on the terrestrially based ^{14}C timescale. Pollen and SSTs both show a similar deglacial pattern, but climatic changes appear to occur about 2000 years earlier in the pollen record. This early response in the pollen record is too large to be an artifact of differences in chronologies applied to the records, suggesting that the atmosphere may have responded earlier to deglacial forcing than the ocean. (c) Timing of the summer SST record in core H214 using the terrestrial (black) and marine (grey) timescale. The ^{14}C timescale is compared to the calendar timescale. The time of the Antarctic cold reversal is shown in the grey bar between 14 and 12.5 cal kyr B.P. [*Jouzel et al.*, 2001]. Events in the terrestrially based ^{14}C timescale are offset by ~500 years from the marine ^{14}C timescale between ~14.0 and 9.0 ka. Cooling in core H214 appears to be coincident (marine timing) or to lead (terrestrial timing) cooling in Antarctica depending on which timescale is used. The ^{14}C ages were converted to calendar ages based on terrestrial tephra dates using the CALIB 5 program (M. Stuiver et al., CALIB 5.0, WWW program and documentation, http://calib.qub.ac.uk/calib/, 2005). Tephra abbreviations follow *Froggatt and Lowe* [1990]: Ma, Mamaku; Rm, Rotoma; Wh, Waiohau; and Rw, Rerewhakaaitu.

that the atmosphere responded faster than the ocean to deglacial forcing in the subtropical southwest Pacific.

[29] The terrestrially based ^{14}C chronology for H214 indicates SSTs warmed by 2°C between ~14.2 and 12.2 ka compared to between ~14.5 and 12.85 ka in the marine ^{14}C chronology (Figure 6c). The subsequent cold interval is centered on ~11.0 ka in the terrestrially based ^{14}C chronology whereas the marine ^{14}C timescale suggests it is centered on 11.5 ka. The temperature maximum observed in the summer SST record is centered on ~10.0 ka in the terrestrially based ^{14}C timescale versus ~10.5 ka for the marine ^{14}C chronology. In general, the terrestrially based ^{14}C chronology suggests that deglacial climatic changes occurred 300–700 years later than implied by the marine ^{14}C timescale. There are fewer control points in the terrestrially based ^{14}C timescale and the effect on the chronology in H214 of different apparent sedimentation rates is only pronounced above the uppermost tephra and below the lowermost tephra where marine and terrestrial ages differ by up to 1500 years. Importantly, in the middle section of the core where the signals of the two climatic proxies diverge, there is better chronologic and stratigraphic control. Here marine and terrestrially based ^{14}C ages typically differ by less than 700 years.

[30] Although the differences between the marine and terrestrially based timescales in H214 appear small, the interhemispheric and ocean-atmospheric leads and lags that we are trying to identify are often equally small and therefore may be dependent on chronologies applied to deglacial climate records. This dependency emphasizes the importance of comparing deglacial records on a common timescale and the need to better establish past changes in surface reservoir ages.

4.3. Younger Dryas or Antarctic Cold Reversal in the Subtropics?

[31] In both planktonic foraminiferal $\delta^{18}O$ records a brief enrichment occurs at ~9.5 ka (Figure 4) about 700 years after the time of the Younger Dryas which has been radiocarbon dated in the marine record at ~10.2 ka [e.g., *Fairbanks*, 1989]. This shift is not reflected in the assemblage SST record, which shows no evidence for cooling at

that time (Figure 5). However, it is evident in the benthic as well as the planktonic isotope records. In all the isotopic records, the signal is small and short-lived; in the *G. ruber* record it is only one point and may reflect noise in the record. Nonetheless, if the signal is real, it may reflect global changes in $\delta^{18}O$ driven by changes in deglacial meltwater inputs associated with the Younger Dryas that are not reflected in temperature. The 700 year lag would suggest transport of this signal through thermohaline circulation, rather than atmospheric transport [*Anderson and Thunell*, 1993]. Either way, the signal is very slight. This small shift in the isotope signal is not seen in other cores from the area, but this may be a result of the higher resolution of the record in H214 [*Nelson et al.*, 2000; *Newnham et al.*, 2003; *Wright*, 1983].

[32] On the South Island of New Zealand, an advance of the Franz Josef Glacier (43°27′S, 170°10′E), which formed the Waiho Loop terminal moraine, has been radiocarbon dated at ~11.05 ka [*Denton and Hendy*, 1994] and was correlated with the Younger Dryas. However, the timing of the deposition of the Waiho Loop terminal moraine has been challenged, as other studies [*Mercer*, 1988] have radiocarbon dated the advance at ~11.6 ka and interpretation of the glacial advance remains controversial [*Fitzsimons*, 1997; *McGlone*, 1995]. Since that study, evidence for and dating of the Antarctic cold reversal has improved [*Jouzel et al.*, 1987, 1992, 1995, 2001]. Both radiocarbon dates for the Waiho Loop fall within the radiocarbon age range of the ACR implying that the advance may correlate with the ACR rather than the Younger Dryas.

[33] The strong and persistent 2°–3°C cooling spanning the interval from 12–5 to 11.0 ka in H214 is clearly too early to be associated with the Younger Dryas. Instead the interval, centered on ~11.5 ka, appears to correlate better with the Antarctic Cold Reversal (Figure 6c). The ACR occurs between 14.0 and 12.5 cal kyr B.P. in Antarctic ice cores [*Blunier et al.*, 1997; *Jouzel et al.*, 1987, 1992, 1995, 2001]. We have converted the ^{14}C timescale in H214 based on terrestrial dates to calendar years using the CALIB program (M. Stuiver et al., CALIB 5.0, WWW program and documentation, 2005, available at http://calib.qub.ac.uk/calib/) and the SST cooling appears synchronous with the ACR. Using the terrestrial ^{14}C timescale, the cooling in H214 begins before the ACR. Thus, by either timescale, this cooling is better correlated with the ACR than the Younger Dryas. The ACR has also been documented in high-resolution marine records from the Southern Ocean [*Labracherie et al.*, 1989; *Pahnke et al.*, 2003]. Notably, one of these sites is in the sub-Antarctic in the New Zealand region. The strong signal in H214 suggests that the cooling associated with the ACR also extended into the Southern Hemisphere subtropics in the southwest Pacific.

5. Conclusions

[34] Core H214 from the Bay of Plenty provides a centennial-scale deglacial record for the subtropical southwest Pacific. A marine ^{14}C chronology was based on several AMS ^{14}C dates on monospecific planktonic foraminiferal samples. The presence of four tephras in the core, which have been correlated to ^{14}C dated tephras on land, enabled the development of a marine record calibrated to the terrestrial ^{14}C timescale for the deglacial record, thus avoiding the uncertainty of the marine radiocarbon reservoir correction.

[35] 1. Foraminiferal assemblage estimates of SST in the late glaciation ~16.5 ka were 15.6°C in winter and 20.3°C in summer. Both assemblage-based SST estimates and isotopic estimates indicate that Bay of Plenty temperatures at the end of the glaciation at ~16.5 ka were 1.5°–2.8°C colder than today.

[36] 2. At the beginning of the Holocene a temperature maximum, 3°C warmer than modern and lasting over 1000 years, was centered on ~10.5 ka in H214. The temperature maximum appears coincident with the early Holocene temperature maximum or "climatic optimum" observed in the high latitudes of the Southern Hemisphere.

[37] 3. In the Bay of Plenty, SST changes lag changes in pollen abundances by ~2 kyr implying that the ocean responded more slowly to deglacial forcing than the atmosphere in the subtropical southwest Pacific.

[38] 4. The deglacial warming associated with the last deglaciation was interrupted by a ~2°C cooling centered on ~11.5 ka. This cooling appears coincident with the Antarctic Cold Reversal.

[39] 5. The temperature records in H214 indicate that cooling associated with the ACR and warming in the early Holocene that have been observed in the high-latitude Southern Hemisphere extended into the Southern Hemisphere subtropics, at least in the southwest Pacific. This implies that the occurrence of the ACR and the early Holocene temperature maximum may have been much more widespread than previously thought.

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W. R. Howard, Cooperative Research Centre for Antarctic and Southern Ocean Environment, GPO Box 252-80, Hobart, Tasmania 7001, Australia.

C. R. Samson, Department of Zoology, University of Tasmania, GPO Box 252-80, Hobart, Tasmania 7001, Australia.

E. L. Sikes, Institute of Marine and Coastal Sciences, Rutgers, The State University of New Jersey, 71 Dudley Road, New Brunswick, NJ 08901-8521, USA. (sikes@imcs.rutgers.edu)