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Radiocarbon age anomaly at intermediate water depth in the Pacific Ocean during the last deglaciation

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[1] Benthic and planktonic ^{14}C ages are presented for the last glacial termination from marine sediment core VM21-30 from 617 m in the eastern equatorial Pacific. The benthic-planktonic ^{14}C age differences in the core increased to more than 6000 years between Heinrich 1 time and the end of the Younger Dryas period. Several replicated ^{14}C ages on different benthic and planktonic species from the same samples within the deglacial section of the core indicate a minimal amount of bioturbation. Scanning electron microscopy reveals no evidence of calcite alteration or contamination. The oxygen isotope stratigraphy of planktonic and benthic foraminifera does not indicate anomalously old (glacial age) values, and there is no evidence of a large negative stable carbon isotope excursion in benthic foraminifera that would indicate input of old carbon from dissociated methane. It appears, therefore, that the benthic ^{14}C excursion in this core is not an artifact of diagenesis, bioturbation, or a pulse of methane. A benthic $\Delta^{14}C$ stratigraphy reconstructed from the ^{14}C ages from the deglacial section of VM21-30 appears to match that of Baja margin core MV99-MC19/GC31/PC08 (705 m), but the magnitude of the low-^{14}C excursion is much larger in the VM21-30 record. This would seem to imply that the VM21-30 core was closer to the source of ^{14}C-depleted waters during the deglaciation, but the source of this CO_2 remains elusive.

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1. Introduction

[2] In a recent paper *Marchitto et al.* [2007] presented evidence for a two-part excursion in radiocarbon activity ($\Delta^{14}C$) at 705 m water depth in the northeastern Pacific Ocean during glacial termination I. These authors demonstrated a roughly 200‰ drop in $\Delta^{14}C$ between ~17 and 15 ka B.P. and another equivalent negative excursion between ~14 ka B.P and 11 ka B.P. The ^{14}C excursions were documented from ^{14}C measurements of benthic foraminiferal calcite extracted from a sediment core whose chronostratigraphy could be tied independently of ^{14}C to the GISP II ice core record (Figure 1).

[3] *Marchitto et al.* [2007] argued that large $\Delta^{14}C$ excursions spanning the time between the Heinrich 1 (H1) and the Younger Dryas (YD) events are indicative of a change in ventilation dynamics within the Pacific during the last deglaciation that released ^{14}C depleted CO_2 from a cold, saline deep water mass [*Adkins et al.*, 2002a, 2002b] that had been isolated from the atmosphere for thousands of years during the glacial [*Keeling and Stephens*, 2001; *Sigman and Boyle*, 2000; *Toggweiler*, 1999, 2008]. They speculated that this renewed ventilation was brought about by sea ice retreat and wind pattern changes around Antarctica, and that it led to ventilation of previously sequestered CO_2 through the Southern Ocean and to subsequent transport of a low-$\Delta^{14}C$ signal into the ocean interior via Antarctic Intermediate Water/sub-Antarctic Mode Water.

2. Background

[4] The database of ^{14}C ages from fossil carbonate precipitated in the ocean's surface water and in speleothems document a systematic and long-term decrease in radiocarbon concentrations ($\Delta^{14}C$) from the last glacial period to the recent Holocene [*Beck et al.*, 2001; *Chiu et al.*, 2007; *Fairbanks et al.*, 2005; *Hughen et al.*, 2004; *Muscheler et al.*, 2005; *Voelker et al.*, 2000] (Figure 1). The magnitude and duration of the $\Delta^{14}C$ changes spanning the past 30,000 years implies the production of ^{14}C in the atmosphere has not remained constant and that the partitioning of $^{14}CO_2$ between reservoirs has varied, The ^{14}C records also show several shorter-term excursions in $\Delta^{14}C$ during the past 30,000 years that are not found in archives of other cosmogenic isotopes and thus do not appear to be due to production changes [*Muscheler et al.*, 2005]. These shorter-term excursions imply that the variable uptake and release of carbon from the ocean had an important influence on the ^{14}C content of the atmosphere.

[5] Among the most notable of the $\Delta^{14}C$ excursions during the past 30,000 years was a ~50‰ rise and subsequent 190‰ drop in atmospheric and surface ocean $\Delta^{14}C$

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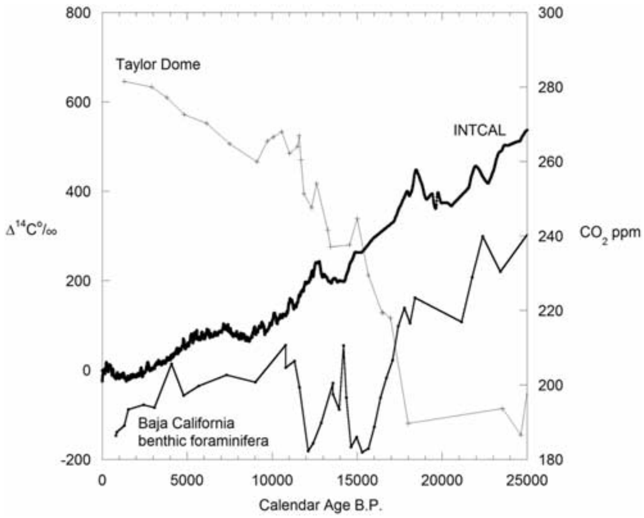

Figure 1. The INCAL $\Delta^{14}C$ record based on tree rings, varves, and corals [*Reimer et al.*, 2004] and Baja margin [*Marchitto et al.*, 2007] compared to the deglacial rise in atmospheric CO_2 from Taylor Dome ice core [*Smith et al.*, 1999].

between 17.5 and 14.5 ka B.P. (Figure 1), coeval with a rise in atmospheric concentrations [*Monnin et al.*, 2001] that points to a release of CO_2 to the atmosphere from a ^{14}C-depleted reservoir. The deglacial rise in atmospheric CO_2 stratigraphy began to rise between 17 ka B.P. and 18 ka B.P., rising by 40 ppmv by 14.5 ka B.P. [*Monnin et al.*, 2001]. The precise timing of the initial increase in CO_2 at the end of the last glacial has not been firmly established because of uncertainties in assigning gas ages to CO_2 records [*Blunier and Brook*, 2001; *EPICA community members*, 2004; *Monnin et al.*, 2001]. In the Dome C ice core, the concentrations of both CO_2 and CH_4 begin to increase above their glacial values at 525 m. In the EDC1 [*EPICA community members*, 2004] time scale this depth has a corresponding gas age of 17 ka B.P.. However, by synchronizing CH_4 records from the Antarctic Dome C core and the Greenland GISP ice core [*Blunier and Brook*, 2001] and adopting the GISPII gas age model, the initial CO_2 rise would correspond to $\sim$18 ka B.P. as originally presented by *Smith et al.* [1999] for the Taylor Dome core. In reviewing the ocean and atmospheric changes at the end of the last glacial *Denton et al.* [2006] and *Broecker and Barker* [2007] referred to the time between 17.5 and 14.5 ka B.P. as the "mystery interval" (MI) because of the wide spread ocean and atmospheric changes that accompanied the enigmatic $\Delta^{14}C$ shift. The MI was a time of massive iceberg discharge in the North Atlantic (Heinrich 1 event) [*Bond et al.*, 1993; *Bond and Lotti*, 1995] and of weakened North Atlantic Deep Water overturning circulation [*McManus et al.*, 2004]. But since the radiocarbon excursion during the MI involved a large drop in $\Delta^{14}C$, it appears to have been caused by factors other than a reduction in North Atlantic overturning circulation itself since this would be expected to have reduced the uptake of $^{14}CO_2$ from the atmosphere.

[6] *Broecker and Barker* [2007] pointed out in their review of available data for the period between 17.5 and 14.5 ka B.P. that in the absence of a sustained release of radiocarbon-dead CO_2 from either volcanism or from a methane outburst, neither of which is evident in ice core records, and with no evidence of dramatic ^{14}C production changes [*Muscheler et al.*, 2005], the most likely source of radiocarbon-depleted CO_2 would have been from the deep sea. But this would require there to have been a significant increase in the age of deep waters during the glacial, perhaps by as much as several thousand years beyond the age of the oldest water masses in the deep ocean today. The absolute size and distribution of such ^{14}C-depleted water is not easily estimated because so few deep sea ^{14}C records currently exist that can be used to reconstruct ^{14}C gradients within the glacial ocean.

[7] There are a number of reasons to consider that an increase in ventilation age could have occurred during the last ice age. Pore water chlorinity data indicate bottom waters in the glacial ocean were more highly saline and more strongly stratified, which would have contributed to reduced ventilation rates [*Adkins et al.*, 2002a]. In the North Atlantic, $\Delta^{14}C$ values in benthic foraminifera from the Last Glacial Maximum and from H1 time in cores collected from below 3000 m are as much as 250‰ lower than modern values [*Robinson et al.*, 2005], implying there was greater proportion of older, more ^{14}C-depleted southern sourced waters in the North Atlantic at times when NADW was reduced. While these ^{14}C results and the evidence of higher deep water salinities could imply a more sluggish circulation, to date there is no evidence from Pacific sites at depths below 2000 m documenting deep waters that were as much as 6000 years old during the last glacial [*Broecker and Barker*, 2007]. Where then would the ^{14}C-depleted waters come from that ventilated the thermocline depths of the eastern North Pacific during the last glacial termination [*Marchitto et al.*, 2007]?

[8] In the present study we have used radiocarbon ages of planktonic and benthic foraminifera to estimate the ^{14}C age of both surface and intermediate waters in the eastern equatorial Pacific across the Mystery Interval in a further effort to evaluate the extent of ^{14}C-depleted waters within the Pacific during the deglaciation. New ^{14}C data is presented for core VM21-30 (Table 1), a core collected from 617 m water depth within the upwelling regime to the east of the Galapagos Islands (Figure 2). We targeted VM21-30 for this study for two reasons. First, we wished to further test whether ^{14}C-depleted waters could be identified at a site located south of the equator where today most of the intermediate waters are ventilated in the Southern Ocean [*McCreary and Lu*, 2001]. Second, because this site lies within the upwelling region of the eastern tropical Pacific, ^{14}C-depleted CO_2 may have ventilated through the upwelling system and would be apparent in the ^{14}C ages of planktonic foraminifera. The sedimentation rate of VM21-30 [*Koutavas et al.*, 2002; *Koutavas and Lynch-Stieglitz*, 2003] is somewhat higher than other open ocean sites, providing some assurance that bioturbation has not completely obscured a record of millennial-scale $\Delta^{14}C$ change. Earlier, *Broecker et al.* [2004] measured very large planktonic-benthic age

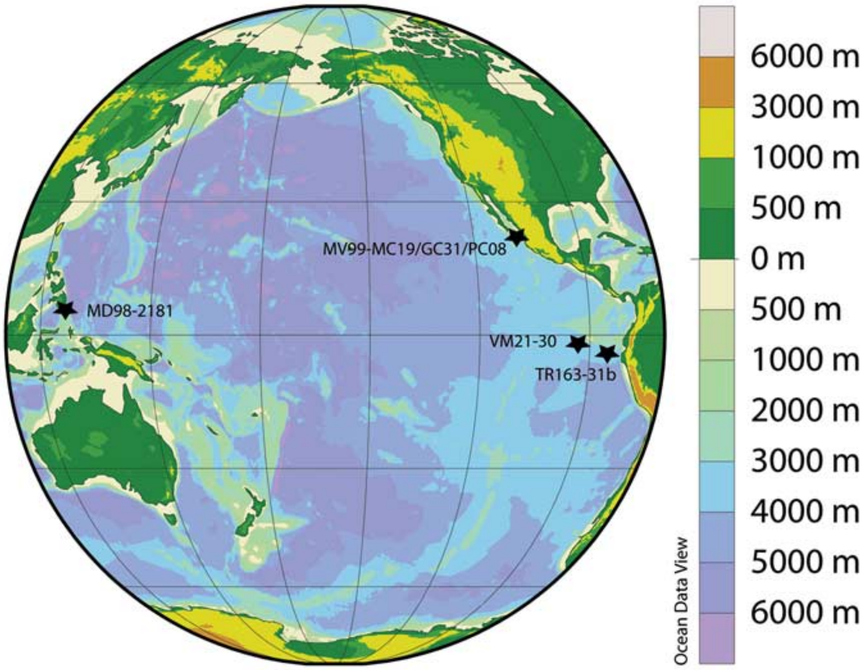

Figure 2. Location map of marine core described in the text. Colors represent bathymetry and topography.

differences from a sample taken at 275–285 cm in this core, but chose not to incorporate the planktonic and benthic ages from this core in their survey of Pacific ^{14}C ventilation ages because of discrepancies in ages for mixed benthics and *Uvigerina*. However, taking the younger of their two VM21-30 benthic ages gives a benthic-planktonic age difference of 3620 years, substantially older than the ~800–1000 year difference that is observed today at this location [*Key et al.*, 2004]. A larger data set of ^{14}C data from VM21-30 is presented here and is compared with the ^{14}C record from the Baja margin [*Marchitto et al.*, 2007].

3. Sample Preparations, Age Models, and $\Delta^{14}C$ Chronologies

[9] Foraminiferal samples for ^{14}C dating were picked from the >250 μm size fraction of VM21-30 core sediments that had been disaggregated in tap water and dried at low temperature. Picked foraminifera were sonified briefly three times in deionized water, removing the supernatant after each rinse. This step was followed by a sonification rinse in methanol and then DI water, followed by a rinse in 0.15% H_20_2 in 0.1 M NaOH at 90°C for 5 min and briefly sonicated twice. After a final rinse in DI water and second rinse in methanol the samples were transferred to a drying oven maintained at 50°C. The ^{14}C samples were typically between 2 and 5 mg, although several of the benthic samples were smaller because of lower abundances. AMS dating was carried out at the WHOI and UCI AMS laboratories (Table 2). Samples measured at UCI were leached 10% with dilute HCl immediately prior to hydrolysis, whereas WHOI samples were not leached. Geological calcite samples were measured at both laboratories to evaluate size-dependent sample processing blanks. Quoted uncertainties for both laboratories are based on scatter of multiple measurements as well as counting statistics, and include contributions from normalization to radiocarbon standards and for background subtraction. Three planktonic and one benthic sample measured at UCI underwent two-step dissolution, with independent ^{14}C measurements of the CO_2 from the two fractions (Table 2).

[10] To develop a $\Delta^{14}C$ stratigraphy for the VM21-30 core through the mystery interval we use several prominent $\delta^{18}O$ and $\delta^{13}C$ datums that are well characterized in the foraminiferal isotope stratigraphy through the last deglaciation. These include the negative shift in marine carbonate $\delta^{18}O$ between 14.5 and 12.3 ka B.P. that marked meltwater event 1A, [*Fairbanks*, 1989; *Peltier*, 2005] and the 0.5‰ $\delta^{13}C$ minimum in atmospheric $^{13}CO_2$ between ~18–11 ka B.P. documented in the Taylor Dome Ice core [*Smith et al.*, 1999] (Figure 3). A corresponding deglacial $\delta^{13}C$ excursion is also evident in the planktonic foraminifera isotope stratigraphies of tropical/subtropical Pacific cores [*Bostock et al.*, 2004; *Spero and Lea*, 2002], allowing for direct correlation between the ice core and tropical marine records (Figure 3 and Table 3). We take the beginning and the end of the deglacial $\delta^{13}C$ excursion in the Taylor Dome Ice Core stratigraphy [*Smith et al.*, 1999] as chronologic datums and use these ages to estimate the sediment accumulation rate

Table 1. Core Locations

Core	Latitude	Longitude	Water Depth (m)
VM21-30	1°13′S	89°41′W	617
MV99-MC19/GC31/PC08	23.5°N	111.6°W	705

Table 2. The ^{14}C Results for VM21-30

Sample Depth (cm)	Weight (mg)	Benthics	Planktonics	^{14}C (years B.P.)	^{14}C Error (years)	Accession Number[a]
0[b]			*Globigerinoides dutertrei*	2,190	45	OS-20675
12	6.13		*Globigerinoides ruber*	1,865	15	UCIAMS-45224
12	6.37		*Globigerinoides sacculifer*	1,875	15	UCIAMS-45225
12	2.93	*Uvigerina*		3,070	20	UCIAMS-45223
27	5.90		*Globigerinoides ruber*	4,270	20	UCIAMS-45227
27	5.84		*Globigerinoides sacculifer*	3,755	20	UCIAMS-45228
27	3.25	*Uvigerina*		4,790	20	UCIAMS-45226
55[b]			*Globigerina dutertrei*	5,620	55	OS-30482
66	3.90	*Uvigerina*		7,545	20	UCIAMS-45229
66	6.35		*Globigerinoides ruber*	6,760	20	UCIAMS-45230
66	3.78		*Globigerinoides sacculifer*	6,335	20	UCIAMS-45231
107–108	2.20	*Uvigerina*		11,850	280	OS-38610
107–108	6.61		*Globigerinoides sacculifer*	8,590	45	OS-38368
107–108	5.44		*Globorotalia menardii*	8,500	20	UCIAMS-39277
114–115	6.09		*Globigerinoides sacculifer*	8,690	55	OS-38375
114–115	5.24		*Globorotalia menardii*	8,525	30	UCIAMS-39278
114–115	1.81	*Uvigerina*		11,750	230	OS-38612
120[b]			*Globigerina dutertrei*	9,830	55	OS-20674
139	3.78		*Globigerinoides sacculifer*	10,050	50	OS-38376
139	3.45		*Globorotalia menardii*	10,090	30	UCIAMS-39279
139	1.18	*Uvigerina*		15,350	360	OS-38620
143	5.77		*Globigerinoides sacculifer*	10,150	55	OS-38370
143	4.84		*Globorotalia menardii*	9,830	20	UCIAMS-39280
143	1.74	*Uvigerina*		16,550	390	OS-38614
150–151	4.35		*Globigerinoides ruber*	11,770	40	UCIAMS-43159
150–151	3.06		*Globigerinoides sacculifer*	10,810	45	UCIAMS-43160
150–151	1.83	*Uvigerina*		17,460	170	UCIAMS-43158
151[b]			*Globigerina dutertrei*	11,250	50	OS-20676
161[b]			*Globigerina dutertrei*	13,400	75	OS-21890
170[b]			*Globigerina dutertrei*	11,750	90	OS-30483
170–171	3.81		*Globigerinoides ruber*	13,590	45	UCIAMS-43162
170–171	2.19		*Globigerinoides sacculifer*	12,680	70	UCIAMS-43163
170–171	2.62	*Uvigerina*		19,730	150	UCIAMS-43161
177	2.17		*Globigerinoides sacculifer*	12,000	130	UCIAMS-38304
177	3.98		*Globorotalia menardii*	11,950	25	UCIAMS-39281
177	1.34	*Oridorsalis*		19,950	150	UCIAMS-39282
177	2.22	*Uvigerina*		21,100	560	OS-38609
181.5	2.81		*Globigerinoides sacculifer*	12,000	55	OS-38408
181.5	2.50		*Globorotalia menardii*	12,405	40	UCIAMS-39283
181.5	1.29	*Uvigerina*		18,650	380	OS-38619
184[b]			*Globigerina dutertrei*	12,250	70	OS-30484
201–202	2.80		*Globigerinoides ruber*	14,540	45	UCIAMS-46970
201–202	3.90		*Globigerinoides sacculifer*	14,020	35	UCIAMS-46971
201–202	2.64	*Uvigerina*		20,320	80	UCIAMS-46969
211	5.03		*Globigerina dutertrei*	15,750	40	UCIAMS-46973
211	2.63		*Globorotalia menardii*	13,290	35	UCIAMS-46974
211	2.49	*Uvigerina*		20,970	90	UCIAMS-46972
215[b]			*Globigerina dutertrei*	13,400	90	OS-30485
216	1.43	Mixed benthic		16,470	80	UCIAMS-46975
221	2.82		*Globigerinoides ruber*	17,015	50	UCIAMS-46977
221	2.71		*Globigerinoides sacculifer*	14,445	45	UCIAMS-46978
221	3.32	*Uvigerina*		20,600	80	UCIAMS-46976
225[c]	2.20		*Globigerina dutertrei* 1	15,650	70	UCIAMS-47875
225[c]	2.50		*Globigerina dutertrei* 2	16,520	80	UCIAMS-47876
225	1.50	Mixed benthic		21,260	150	UCIAMS-47864
236[c]	2.70		*Globigerina dutertrei* 1	17,550	80	UCIAMS-47877
236[c]	2.70		*Globigerina dutertrei* 2	17,430	80	UCIAMS-47878
236	1.50	Mixed benthic		19,120	110	UCIAMS-47865
240	0.90		*Globigerinoides sacculifer*	14,830	70	
241	1.80		*Globigerina dutertrei* 1	16,710	90	UCIAMS-47879
241	2.30		*Globigerina dutertrei* 2	17,140	80	UCIAMS-47880
241	2.20	*Uvigerina* 1		21,580	140	UCIAMS-47881
241	1.40	*Uvigerina* 2		21,380	180	UCIAMS-47882
251	1.20		*Globigerinoides sacculifer*	18,560	100	UCIAMS-47867
250[b]			*Globigerina dutertrei*	16,800	80	OS-20680
251	4.67	Mixed benthic		22,760	90	UCIAMS-46979
268–269	3.58		*Globigerinoides ruber*	20,510	80	UCIAMS-39284
268–269	4.16	*Oridorsalis*		24,440	80	UCIAMS-39289
268–269	7.69	*Uvigerina*		24,430	90	UCIAMS-39290

Table 2. (continued)

Sample Depth (cm)	Weight (mg)	Benthics	Planktonics	^{14}C (years B.P.)	^{14}C Error (years)	Accession Number[a]
278–280	5.68		*Globigerinoides ruber*	21,330	80	UCIAMS-39291
278–280	8.21	*Oridorsalis*		26,650	110	UCIAMS-39292
278–280	7.85	*Uvigerina*		23,180	70	UCIAMS-39293
320[b]			*Globigerina dutertrei*	25,100	150	OS-30486
400[b]			*Globigerina dutertrei*	26,900	150	OS-15720

[a]UCI results obtained in the present study have been corrected for isotopic fractionation according to the conventions of *Stuiver and Polach* [1977], with $\delta^{13}C$ values measured on prepared graphite using the AMS spectrometer. These can differ from $\delta^{13}C$ of the original material, if fractionation occurred during sample graphitization or the AMS measurement, and are not shown.

[b]From [*Koutavas and Lynch-Stieglitz*, 2003].

[c]Two-step dissolution.

(age model) through the deglacial section of VM21-30 (Table 3). We note that the chronology of these datums in the Taylor Dome Ice core has uncertainties of several centuries [*Monnin et al.*, 2001]. We also recognize that the Taylor Dome age model as presented by *Grootes et al.* [2001] and as presented by *Smith et al.* [1999] differ in the glacial section of the core, prior to ~18 ka B.P. By using the ice core datums to align the marine and ice core stratigraphies through the deglacial sections we have attempted to circumvent potential error that would arise from varying marine ^{14}C reservoir ages in the eastern tropical Pacific. In the Holocene and LGM sections however, there are no such datums that tie the ice and marine core stratigraphies together. The VM21-30 age model for the Holocene and for the LGM interval was developed by applying conventional ^{14}C calendar year calibrations to the planktonic ^{14}C ages INTCAL04 [*Reimer et al.*, 2004] with a 500 year reservoir age correction. We acknowledge that this reservoir correction introduces an unknown level of age uncertainty. The magnitude of this uncertainty cannot be fully assessed but we compare the resulting D14C stratigraphy for VM21-30 to that of Cariaco Basin [*Fairbanks et al.*, 2005; *Hughen et al.*, 2004] (Figure 4). The comparison suggests that the constant reservoir age assumption is not too far off because in the LGM and in the Holocene the two records are reasonably similar. In the deglacial section of the VM21-30

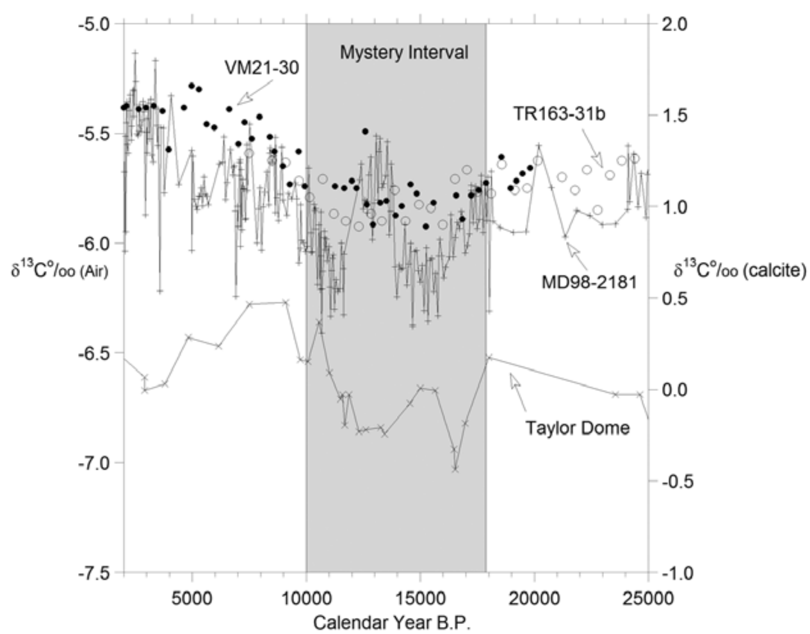

Figure 3. The deglacial $\delta^{13}C_{(VSMOW)}$ stratigraphy of tropical Pacific planktonic foraminifera from eastern equatorial cores VM21-30 (*Globigerinoides ruber* [*Koutavas and Lynch-Stieglitz*, 2003]) and TR163-31b (*Globigerinoides dutertrei* [*Patrick and Thunell*, 1997]) and western tropical Pacific core MD98-2181 (*Globigerinoides ruber* [*Stott et al.*, 2004]) with deglacial $\delta^{13}C$ stratigraphy for atmospheric CO_2 from Taylor Dome ice core [*Smith et al.*, 1999]. The mystery interval is marked in gray. The beginning of the deglacial $\delta^{13}C$ excursion is less certain in the ice core because of the resolution of the measurements. However, each of the marine records documents a $\delta^{13}C$ excursion during the early part of the deglaciation that is similar in magnitude to the atmospheric signal and is characterized by two minima, one centered at 16–17 ka B.P. and the second centered between 11.5 and 13.5 ka B.P. The MD98-2181 core has a disturbed interval between 961 and 941 cm [*Stott*, 2007], so the precise position of this second $\delta^{13}C$ minimum is less certain.

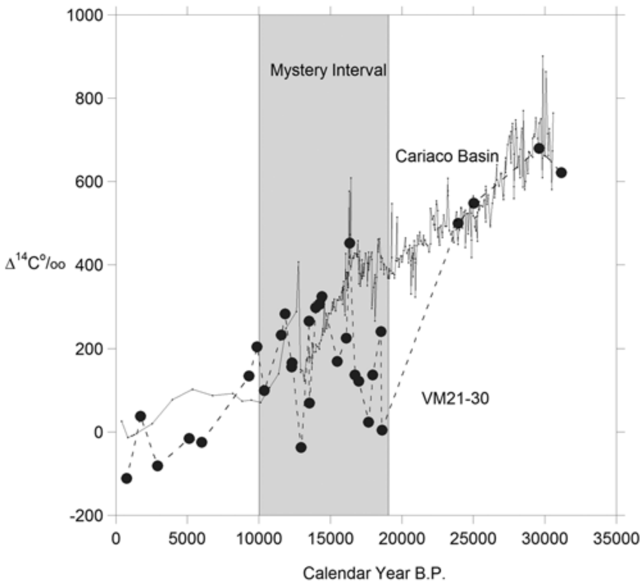

Figure 4. The planktonic $\Delta^{14}C$ stratigraphy of VM21-30 (dashed) in comparison with the planktonic $\Delta^{14}C$ stratigraphy from Cariaco Basin (solid) [*Hughen et al.*, 2006].

record there is a marked offset relative to the Cariaco Basin record (Figure 4). The VM21-30 values are markedly lower during the Mystery Interval, coinciding with the large benthic-planktonic ^{14}C differences. One potential implication of this would be that old radiocarbon was being upwelled into the surface waters in the eastern equatorial Pacific during the Mystery Interval.

4. Preservation of Foraminiferal Calcite

[11] We assessed the preservation of VM21-30 foraminifera and the efficiency of our cleaning techniques with scanning electron microscope (SEM) examination both benthic and planktonic foraminiferal tests from several core samples (Figure 5). Examination of the surficial and cross sections of shell calcite of both benthic and planktonic specimens revealed no evidence of alteration or secondary calcite. The results convinced us that the carbonate from the VM21-30 core is well preserved and contains very little contaminating detritus after cleaning.

5. VM21-30 ^{14}C Results

[12] The ^{14}C ages of benthic and planktonic foraminifera are shown in Figure 6 and listed in Table 2. The average sediment accumulation rate based on the planktonic ^{14}C age/depth trend is ~10 cm/ka, relatively high for an open ocean site. Nonetheless, the planktonic ^{14}C ages from the VM21-30 core exhibit a good deal of scatter and this presumably reflects the influence of bioturbation. The down core planktonic ^{14}C ages do not however, exhibit any large deviations or excursions whereas the down core trend of benthic ^{14}C ages from this core does exhibit a distinctive nonlinear trend with depth (Figure 6). At the top of the core the benthic ages are offset from the planktonic ages by an average of 1000 years. Below 100 cm the benthic-planktonic age differences increase significantly. Between 100 and 115 cm the benthic-planktonic ^{14}C age difference is ~3000 years and increases to more than 6000 years at 170 cm. A maximum benthic-planktonic age difference of 8550 years occurs at the 177 cm horizon. Below 177 cm the Δ ages decrease again to roughly 3000 years.

[13] We measured ^{14}C ages of two planktonic and benthic foraminiferal species from several different sample horizons. Benthic samples of *Oridosalis* and *Uvigerina* were measured from samples at 177 cm, 268.5 cm and 279 cm. We also measured two successive leaches of *Uvigerina* from the 241 cm sample. Both benthic species from the 177 cm and 268.5 cm samples returned ages that are indistinguishable from one another whereas the 279 cm sample returned an age for *Oridosalis* that is 3400 years older than *Uvigerina*. The two *Uvigerina* leaches from the 241 cm sample are again, indistinguishable from one another (Table 2). In each of these samples the benthic-planktonic age differences are significantly larger than are the samples from the Holocene portion of this core and as much as 3000 to 5000 years greater than the age difference between surface waters and intermediate waters today in the EEP. At the same time, there are two samples from the deglacial section of the core for which the benthic ^{14}C ages are clearly much younger than the benthic ages from above and below. A mixed benthic sample from 215 cm returned an age of 16,470 years whereas *Uvigerina* from the 211 cm horizon is 20,970 years and the *Uvigerina* from the 221 cm horizon is 20,600 years. At 236 cm the mixed benthic ^{14}C age is also younger than are the samples above and below this horizon (Table 2).

[14] The ^{14}C reproducibility of planktonic foraminiferal species varies among samples (Table 2). From several samples specimens of *Globigerinoides ruber* returned ^{14}C ages that are older than *Globigerinoides sacculifer*. But in samples that exhibit the largest benthic-planktonic age difference (~200–100 cm), the planktonic species have very similar ^{14}C ages. In fact, for the 177 cm sample both *G. sacculifer* and *Globorotalia menardii* returned ^{14}C ages of 12,000 +/− 130 years and 11,950 +/− 25 years, respectively. The benthic samples of *Oridosalis* and *Uvigerina* from this same sample are 19,950 +/− 150 and 21,100 +/− 560, respectively. If such good reproducibility for both the planktonic and the benthic ^{14}C ages is taken to be an indicator of how reliable these ages measurements are, it means that the surface to intermediate water ^{14}C age difference was as much as 8000 years during the last deglaciation. This result would seem inconceivable if it were not for the results of *Marchitto et al.* [2007]. These anomalously large benthic-planktonic age differences are restricted to the deglacial section of the core and coincide at least approximately

Table 3. Calendar Ages of Stable Isotope Datums in Pacific Cores[a]

Core	Top $\delta^{13}C$ Minima (cm)	Meltwater 1A $\delta^{18}O$ Shift (cm)	Base $\delta^{13}C$ Minima (cm)
VM21-30	116	182	250
TR163-31b	30	80	96
MD98-2181	860	1080	1238

[a]Top $\delta^{13}C$ minima, 10 ka B.P.; meltwater 1A $\delta^{18}O$ shift, 14.5 ka B.P.; base $\delta^{13}C$ minima, 17.5 ka B.P.

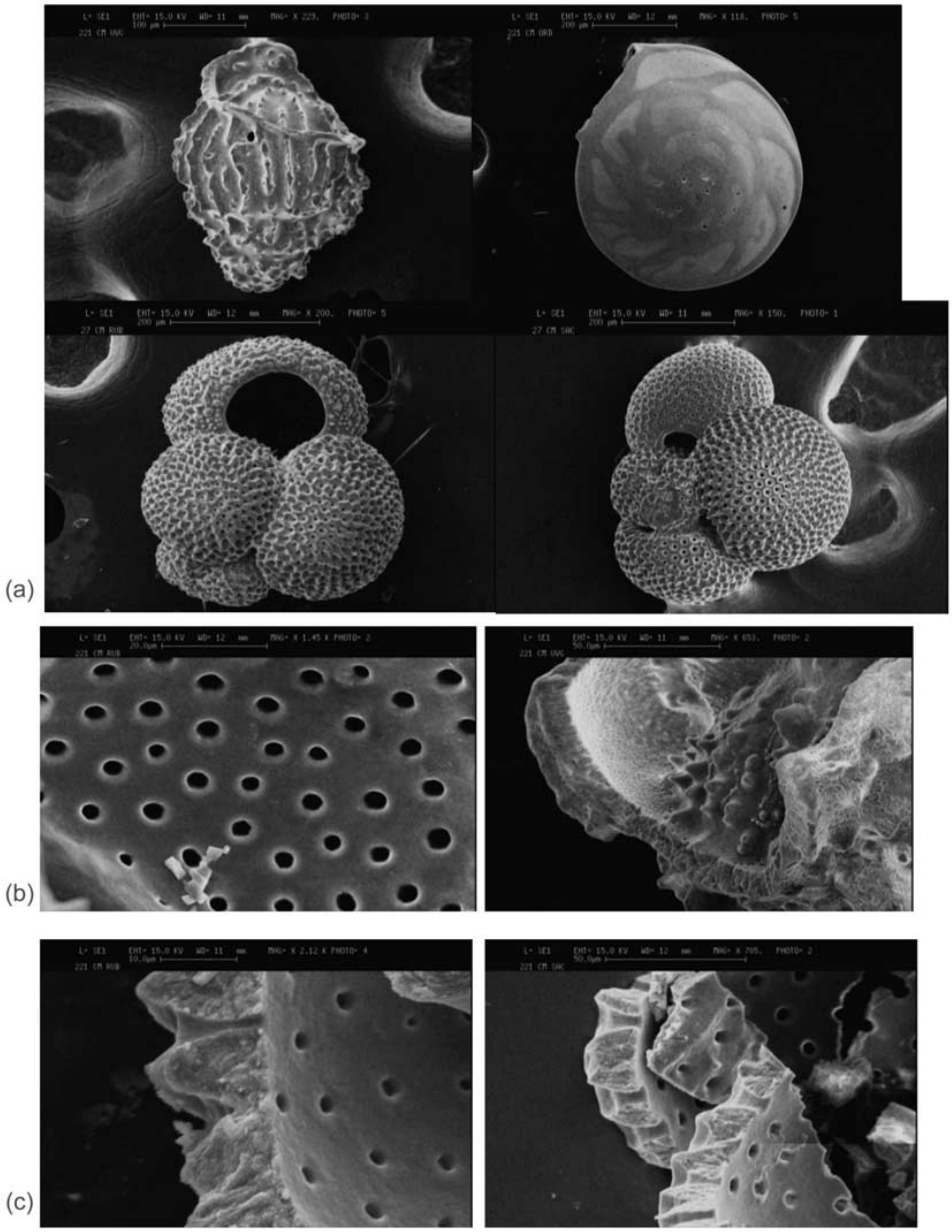

Figure 5. Foraminifera from VM21-30. (a) (clockwise from top left) *Uvigerina* (221 cm), *Oridosalis* (221 cm), *Globigerinoides ruber* (27 cm), and *Globigerinoides sacculifer* (27 cm). (b) (left) Interior of *G. ruber* (221 cm) showing clean and clear pores and wall surface with small rhombs of nonforaminiferal material and (right) interior of *Uvigerina* (221 cm). (c) Cross sections through walls of (left) *G. ruber* (221 cm) and (right) *G. sacculifer* (221 cm). All interior surfaces are clean of detritus, and the pores are open. There is no evidence of diagenesis.

with the timing of the $\Delta^{14}C$ excursion that has been documented from the Baja margin. Furthermore, we see no evidence of a preservation bias between the Holocene and the deglacial samples of this core and the abundances of benthic foraminifera are similar in the Holocene and deglacial sections of the core. There is no marked excursion in the oxygen isotope stratigraphy of *Uvigerina* that would indicate the specimens from the deglacial section have been displaced from older, glacial age horizons (Figure 7). Indeed, the entire benthic glacial Holocene $\delta^{18}O$ stratigraphy agrees

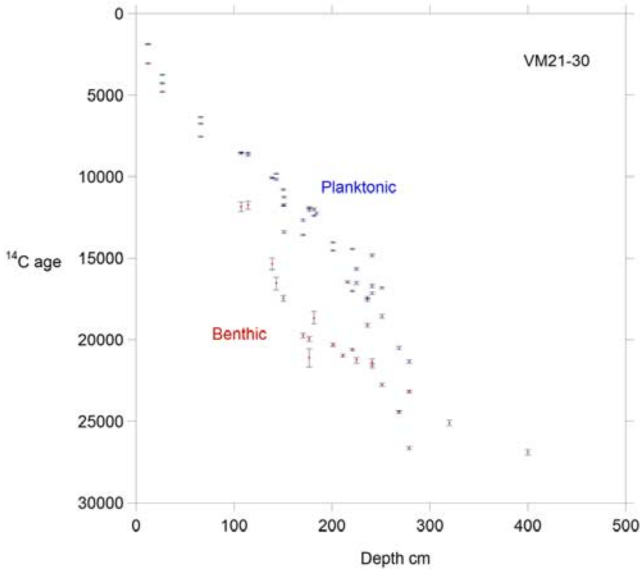

Figure 6. Benthic (red) and planktonic (blue) foraminiferal ^{14}C ages for VM21-30. The ^{14}C ages are plotted with no reservoir age correction.

very well with other Pacific records. Hence, even if the samples that do not have excellent ^{14}C agreement among multiple species are excluded, we are still left with a suite of samples from the deglacial section of VM21-30 for which we find no basis to discount what appears to be a very large benthic-planktonic age differences during the deglaciation.

6. Intermediate Water $\Delta^{14}C$ Anomaly During the Mystery Interval

[15] The available ^{14}C data from the Pacific, including the new data presented here from VM21-30 presents us with a conundrum. On the one hand, the occurrence of a large increase in the benthic to planktonic ^{14}C age difference in the deglacial section of VM21-30 coincides with the well defined excursion that was documented from the Baja margin. On the other hand, the magnitude of the benthic-planktonic age difference in the deglacial section of the VM21-30 core implies that intermediate water ages were many thousands of years older than they are today, and far older than those indicated by the Baja data. Because the benthic glacial Holocene $\delta^{18}O$ stratigraphy agrees with other Pacific records, there is no indication that the deglacial benthic ^{14}C values are an artifact of reworked, glacial age specimens (Figure 7). In fact, the reproducibility among different benthic and planktonic foraminiferal species from several deglacial samples is remarkable for a core with this sediment accumulation rate. Hence, we have to consider whether such old ^{14}C ages are an accurate reflection of intermediate water age during the deglaciation.

[16] One possible explanation for such anomalous deglacial benthic ^{14}C ages could be that there was a pulse of methane injected into intermediate waters of the southeastern Pacific. As *Broecker and Barker* [2007] pointed out, the lack of evidence from ice core records of a large increase in atmospheric methane during the deglacial Mystery Interval rules out any catastrophic releases, but perhaps a lesser injection of clathrate methane into intermediate waters of the south Pacific could have influenced the ^{14}C content of dissolved CO_2 at intermediate water depth in the SE Pacific without affecting atmospheric CH_4 concentrations. However, to have decreased the $\Delta^{14}C$ of the intermediate waters by several hundred per mil during the Mystery Interval would require adding on the order of 3000 μmol/kg additional radiocarbon dead carbon to the $\sim$2200 μmol/kg of dissolved inorganic carbon present at intermediate depths in the southeast Pacific today. Such a large increase in dissolved CO_2 would have produced a calcite dissolution spike and a large negative $\delta^{13}C$ anomaly in benthic foraminiferal calcite since methane is depleted in ^{13}C by as much as 50 to 90‰. There is no evidence for either a dissolution spike or a large negative $\delta^{13}C$ excursion in VM21-30 benthic foraminifera. The benthic $\delta^{13}C$ stratigraphy of this core does exhibit a small, $\sim$0.2‰ decrease between 7 and 11 ka B.P., but these lower values occur after the $\Delta^{14}C$ excursion.

[17] The majority of the $\Delta^{14}C$ excursion that is evident in the intermediate layer of the eastern Pacific occurred during Heinrich event I (Figure 8). Modeling results [*Krebs and Timmermann*, 2007] suggest that a substantial weakening of the AMOC may have led to a reduction of Indonesian Throughflow. This finding has recently been confirmed with the NCAR CCSM3 glacial water hosing simulation (not shown). As demonstrated by idealized layer modeling studies of *McCreary and Lu* [2001], a substantial weakening of the Indonesian Throughflow leads to an intrusion of the North Pacific Intermediate Water (NPIW) into the tropical Pacific intermediate layers. In general, North Pacific Intermediate water is much older than SAMW and AAIW, although not so old as is implied by the benthic ^{14}C anomaly in VM21-30. A switch to a NPIW-dominated regime in the eastern tropical Pacific would immediately explain a substantial aging of the intermediate water masses in the

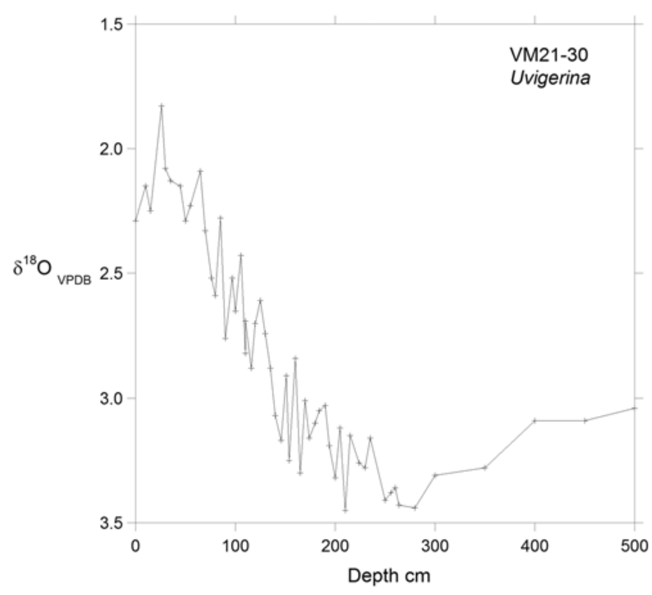

Figure 7. The *Uvigerina* $\delta^{18}O_{VPDB}$ from VM21-30.

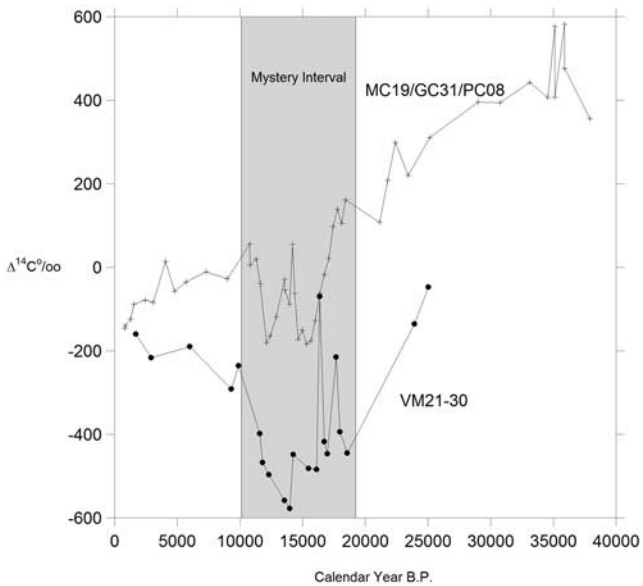

Figure 8. The Δ^{14}C stratigraphies of eastern Pacific cores VM21-30 and MC19/GC31/PC08.

equatorial Pacific. Nonetheless, in order to explain the results presented here in terms of this scenario, the intermediate to deep waters in the North Pacific must have been significantly older prior to Heinrich event I than they are today.

[18] In recent studies of ^{14}C age differences for paired benthic and planktonic foraminifera from ODP887 (northwestern Pacific, 3647 m depth [*Galbraith et al.*, 2007]) and core MR01-K03 from a water depth of 1366 m south of Hokkaido [*Ahagon et al.*, 2003] indicate that LGM intermediate to deep waters were about 1000 years older than during Heinrich I. Neither of these records document water mass ages of up to 6000 years. On the other hand, *Sikes et al.* [2000] did present evidence for ~5500 year old glacial deep water from a core taken east of New Zealand.

7. Conclusions

[19] A number of studies have described the deglacial history of atmospheric CO_2 rise in relation to changing ocean ventilation dynamics, particularly in the Southern Ocean [*de Boer et al.*, 2007; *Keeling and Stephens*, 2001; *Lamy et al.*, 2004; *Marchitto et al.*, 2007; *Pena et al.*, 2008; *Toggweiler et al.*, 1991; *Toggweiler*, 2008]. The results we present for VM21-30 potentially provide an additional line of evidence to support a link between atmospheric CO_2 and ocean ventilation dynamics during deglaciation. The key to verifying the dynamical relationship between atmospheric CO_2 and ocean ventilation change lies in documenting the phasing of events during deglaciation and also identifying the actual reservoir for storage of CO_2 during the glacial. Progress is being made on the first count but the location of CO_2 storage within the ocean during the glacial remains elusive. The very old ^{14}C ages documented in the VM21-30 core would imply that a reservoir of CO_2 was isolated in the ocean for many thousands of years. Around 19–16 ka ago major climatic reorganizations began in both the Northern and Southern Hemispheres. Whereas in the northern hemisphere Heinrich Event I began around 17 ka B.P. and led to the apparent weakening of the AMOC [*McManus et al.*, 2004] and to enhanced ventilation of intermediate to deep layers in the North Pacific [*Gebhardt et al.*, 2008; *Sarnthein et al.*, 2007], there is compelling evidence [*Stott et al.*, 2007, and references therein] that documents earlier warming at higher southern latitudes began around 19–18 ka B.P. This warming can be partly attributed to changes in the austral spring insolation (or equivalently to a lengthening of the summer season) leading to a decrease of austral spring and summer sea ice extent [*Timmermann et al.*, 2009]. Changes near the springtime sea ice margin are likely to have affected temperatures in the main formation sites of AAIW, SAMW and CDW. These temperature changes would have propagated with some delay through the Pacific Ocean [*Stott et al.*, 2007] perhaps contributing to the release of CO_2 from the ocean. But in the absence of evidence for a CO_2-rich, ^{14}C-depleted water masses in the deep Pacific, it remains an open question where the CO_2 was stored during the glacial and what triggered its release during the Mystery Interval.

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