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3 **Radiocarbon Evidence for a Possible Abyssal Front**
4 **Near 3.1 km in the Glacial Equatorial Pacific Ocean**
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22 **ABSTRACT**

23 We investigate the radiocarbon ventilation age in deep equatorial Pacific
24 sediment cores using the difference in conventional ¹⁴C age between coexisting
25 benthic and planktonic foraminifera, and integrate those results with similar
26 data from around the North Pacific Ocean in a reconstruction for the last
27 glaciation (15 to 25 conventional ¹⁴C ka). Most new data from both the
28 Equatorial Pacific and the Emperor Seamounts in the northwestern Pacific
29 come from maxima in abundance of benthic taxa because this strategy reduces
30 the effect of bioturbation. Although there remains considerable scatter in the
31 ventilation age estimates, on average, ventilation ages in the Equatorial Pacific
32 were significantly greater below 3.2 km (~3080 ± 1125 yrs, n=15) than in the
33 depth interval 1.9 to 3.0 km (~1610 ± 250 yrs, n=12). When compared to the
34 average modern seawater Δ¹⁴C profile for the North Pacific, the Equatorial
35 Pacific glacial data suggest an abyssal front located somewhere between 3.0
36 and 3.2 km modern water depth. Above that depth, the data may indicate
37 slightly better ventilation than today, and below that depth, glacial Equatorial
38 Pacific data appear to be as old as last glacial maximum (LGM) deep water
39 ages reported for the deep southern Atlantic. This suggests that a glacial
40 reservoir of aged waters extended throughout the circumpolar Southern Ocean

41 and into the Equatorial Pacific. Renewed ventilation of such a large volume of
42 aged (and, by corollary, carbon-rich) water would help to account for the rise in
43 atmospheric pCO₂ and the fall in Δ¹⁴C as the glaciation drew to a close.

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45 Key Words: radiocarbon; foraminifera; ocean ventilation; Pacific Ocean

46

47 1.0 INTRODUCTION

48 The ocean is the largest reservoir of readily exchangeable carbon on
49 Earth. With ~39,000 gigatonnes of carbon (GTC = 10¹⁵ g C) in the ocean, and
50 ~580 GTC in the recent pre-industrial atmosphere, it is clear that small changes
51 in oceanic carbon storage could cause large changes in the atmospheric CO₂
52 inventory. For this reason, beginning ~25 years ago with ice core observations
53 that the glacial atmosphere had lower CO₂ levels than today (Barnola et al.,
54 1987; Petit et al., 1999), climatologists have looked to the ocean for evidence
55 that its carbon storage was different in the past. Although the surface ocean
56 readily exchanges with the atmosphere if it is not ice covered, the deep ocean
57 exchanges with the atmosphere only at high latitude locations in the North
58 Atlantic and Southern Oceans. Multiple lines of geochemical evidence indicate
59 this exchange was interrupted during cold episodes of the last glacial cycle,
60 either through the stabilizing effects of low salinity at sites of deep-water
61 formation, sea ice cover at those sites, or both (Boyle, 1988; Duplessy et al.,
62 1988; Sigman et al., 2010).

63 Unlike other proxies of deep ocean ventilation, radiocarbon comes with a
64 timescale. Deep ocean ¹⁴C content is set by the atmospheric ¹⁴C activity and the
65 extent of surface ocean equilibration with the atmosphere in the formation
66 region, and decreases quantitatively due to radioactive decay in the ocean
67 interior and may be further influenced by mixing. Available data from solitary
68 corals and foraminifera indicate greater ventilation ages during the last ice age
69 in the North Atlantic (Robinson et al. 2005), the South Atlantic (Skinner et al.,
70 2010), and in the southwest Pacific (Skinner et al. 2015), but the most recent
71 compilations from the Equatorial and North Pacific have failed to identify deep
72 water old enough to account for elevated atmospheric ¹⁴C activity during the
73 last glaciation and the subsequent decline that began about 16.5 ka, at least
74 according to simple mass balance calculations (Broecker, 2004; Broecker and
75 Barker, 2007). For example, Broecker et al. (2004) and Broecker et al. (2008)
76 found no signal of poor ventilation at 2 km – 2.8 km water depth in the west
77 equatorial Pacific (WEP), nor did Broecker and Clark (2010) at 4.4 km in the
78 east equatorial Pacific (EEP). Finally, Broecker and Clark (2011) concluded
79 that the interplay between selective dissolution and bioturbation of planktonic
80 foraminifera (Barker et al., 2007) is too significant to extract meaningful

81 ventilation ages from benthic-planktic age pairs in the deep Pacific. Evidence of
82 significantly older bottom waters than recorded by Broecker and colleagues has
83 been found in cores from intermediate water depths off Baja California
84 (Marchitto et al., 2007), in the Panama Basin (Stott et al., 2009), off the Arabian
85 peninsula (Bryan et al., 2010), in the southwest Pacific (Sikes et al., 2000;
86 Skinner et al., 2015), and in the Southern Ocean (Burke and Robinson, 2012).
87 It has been hypothesized that the deep reservoir mixed with near surface waters
88 in the Southern Ocean and was introduced to Antarctic Intermediate Water
89 (AAIW), yet the signal has not been detected along present-day AAIW
90 pathways located along the western margins of S. America (dePol-Holz et al.,
91 2009) and Africa (Cleroux et al. 2011).

92 There are several metrics for ^{14}C -based ocean ventilation (e.g. Cook and
93 Keigwin, 2015), and of these we focus here on the difference in conventional
94 ^{14}C ages between coexisting benthic and planktonic foraminifera (BF and PF,
95 resectively) which represents the extent of ^{14}C aging or disequilibrium
96 between the surface and deep ocean. Our new ^{14}C results from several cores in
97 the Equatorial Pacific indicate that below ~ 3.2 km, BF-PF ^{14}C ventilation ages
98 were 3080 ± 1125 years during the glacial interval. The contrast of those old
99 deep waters with younger waters above ~ 3.2 km, when compared to the modern
100 $\Delta^{14}\text{C}$ distribution, likely reflects an abyssal front extending from the Southern
101 Ocean into the tropical North Pacific Ocean, although its northern extra-tropical
102 extent remains unclear. In addition to helping explain elevated atmospheric
103 $\Delta^{14}\text{C}$ during the glaciation, subsequent ventilation of these old deep waters
104 likely also influenced the postglacial decline of atmospheric $\Delta^{14}\text{C}$ and the
105 coincident rise in pCO_2 (Marchitto et al., 2007; Anderson et al. 2009; Burke and
106 Robinson, 2012).

107

108 2.0 METHODS

109 This paper presents new data on cores from four locations: the northern
110 Emperor Seamounts in the NWP; the Gulf of California; and the west and the
111 east equatorial Pacific (Table 1). Three of the EEP cores were collected in 1978
112 on R/V Knorr (KNR) cruise 73 by R.P. von Herzen for heat flow studies and
113 were dried out by the time of our sampling. KNR73 piston cores (PCs) 3, 4,
114 and 6 straddle the Equator in a depth transect on the west flank of the East
115 Pacific Rise, from 3.6, 3.7, and 3.8 km water depth, respectively (Figure 1).
116 Core 3PC was studied previously (Boyle and Keigwin 1985) and heavily
117 sampled, so for the present study we sampled the archive half. Samples of three
118 other EEP cores were provided by the curators at Scripps Inst. of Oceanography
119 (PLDS 7G) and Oregon State University (VNTR01-10GC and ME0005 24JC).

120 Samples from WEP free fall core S67 (FFC) 15, were provided by the Hawaii
121 Institute of Geophysics (HIG).

122 Sampling and analytical methods are similar to those reported earlier
123 (Keigwin 2004). At each location we relied on $\delta^{18}\text{O}$ stratigraphy and ^{14}C dating
124 of near-surface dwelling planktonic foraminifera (PF) to identify the glacial
125 interval. As discussed previously, we sought to identify peaks in abundance of
126 benthic foraminifera (BF) for ^{14}C measurement, since the burrowing action of
127 animals on the seafloor tends to reduce the amplitude of abundance maxima but
128 it does not usually create peaks (Broecker et al., 1984). For core ME0005
129 24JC, our sampling of the glacial interval was guided by the chronology of
130 Kienast et al. (2006).

131

132 Because the KNR73 cores were dry and brittle, and not easily sampled
133 without disturbance, we scribed the split surface at 1-cm intervals, removed
134 large sections from the liner, and cut them along the scribe lines. Although this
135 breaks up the sample, large pieces were collected and bagged. Small samples
136 (1-2 g) were removed from the bags for preliminary counting. Intervals of
137 abundant BF were re-sampled at 1-cm spacing to identify the level of peak
138 abundance, and again if necessary until enough BF were recovered for ^{14}C
139 measurement (generally >2 mg CaCO_3). Foraminifera were picked from the
140 fraction >150 μm for both stable isotope and ^{14}C analysis. Where abundance of
141 PF was determined (PCs 3, 4, PLDS 7G, and VNTR01-10GC), samples were
142 microsplit to about 300 individuals and counted (>150 μm fraction). The other
143 EEP cores were still moist and sliced and bagged as above at SIO and OSU, and
144 the WEP free fall core (FFC) was sampled using paleomagnetic cubes by the
145 curator at HIG.

146

147 BF samples were cleaned ultrasonically if necessary, and planktonic
148 samples were cleaned this way only if they were abundant. Smaller samples
149 were put in 1 dram screw cap vials filled completely with distilled water. The
150 vials were shaken to suspend the foraminifera and then quickly hit against the
151 lab bench. This “infrasound” treatment successfully shakes free loosely
152 adhering sediment without too much damage to the shells and is easier to
153 control than ultrasound.

154 Emperor Seamount results and previously published data from shallow
155 locations in the nearby Okhotsk Sea are based on the benthic foram *Uvigerina*
156 and the planktonic *Neogloboquadrina pachyderma* s. because these are the only
157 species present in the glacial interval that were sufficiently abundant for ^{14}C
158 measurement. During warm episodes of deglaciation, ^{14}C dates were on
159 *Globigerina bulloides* and *Uvigerina*. *Uvigerina* was also analyzed from the

160 LGM of DSDP Site 480 samples in the Gulf of California (Keigwin and Jones,
161 1990). In the Equatorial Pacific we dated the PF *Globigerinoides ruber* and
162 *Globigerinoides sacculifer*, and the BF *Nuttalides umbonifera*, *Uvigerina*, and
163 *Cibicidoides*. Low BF abundance and lack of obvious abundance maxima led
164 us to date mixed BF in ME005 24JC, but *Uvigerina* was the most abundant
165 taxon.

166 Accelerator mass spectrometry (AMS) ^{14}C dating was done using routine
167 methods at the National Ocean Sciences AMS (NOSAMS) facility at Woods
168 Hole Oceanographic Institution and at the Univ. of California, Irvine, for
169 samples graphitized at the Univ. of Colorado. All ^{14}C ages are given as the
170 conventional laboratory reported age, without correction for ocean reservoir
171 effects (Stuiver and Polach, 1977). Most stable isotope measurements were
172 made at NOSAMS on a VG Prism mass spectrometer and some were made on a
173 Finigan 252. Where possible, we made analyses on individual specimens of
174 *Uvigerina*, *Cibicidoides*, or *N. umbonifera*.

175

176 3.0 RESULTS

177

178 3.1 Equatorial Pacific.

179 New stable isotope, abundance, and ^{14}C results for Equatorial Pacific
180 cores are summarized in Figure 2 and arranged according to water depth.
181 Stable isotope data are archived at the National Climate Data Center
182 (<http://search.usa.gov/search?affiliate=NCDC&query=paleo>). Where multiple
183 benthic measurements were made at the same level (within 1.0 cm), we report
184 the apparent ventilation age as determined from each benthic species as the
185 difference between the ^{14}C age for that species and the average of the PF ^{14}C
186 ages for that level. All new ^{14}C results and ventilation ages are listed in Table 2.
187 Uncertainty in average PF age is given as the standard error (SE) of the mean of
188 the measurement values or the individual measurement uncertainties summed in
189 quadrature, whichever is larger. Uncertainty of BF-PF ventilation age estimates
190 was calculated as the quadrature sum of the 1σ BF measurement uncertainty
191 and the PF age uncertainty (where only a single PF age is available, the average
192 SE for all PF measurement pairs was used). Where we have AMS dates on
193 multiple BFs from the same stratigraphic level, we calculated ventilation ages
194 for each BF sample separately. Thus, one core depth may provide multiple
195 ventilation age estimates.

196

197

198 For purposes of the present study, we define the glacial interval as
199 between 25 and 15 ^{14}C ka measured in PF. Data are considered to be glacial if

200 one of a pair of PF dates falls within 1σ of the 15-25 ^{14}C ka window. Above
201 ~3.7 km water depth, the glacial interval (shaded interval in Figures 2-4) is
202 found between about 40 and 80 cm, but at greater water depths it appears closer
203 to the core top. As expected, where $\delta^{18}\text{O}$ is maximum $\delta^{13}\text{C}$ is at or near
204 minimum, and the glacial $\delta^{13}\text{C}$ of individual *Cibicidoides* is within the range of
205 previous studies.

206 The benthic fauna changes as a function of water depth, and of the three
207 genera we counted, only *Cibicidoides* are found at all water depths. At all but
208 the deepest site, in the WEP, and the easternmost site (24JC), *Cibicidoides*
209 reach peak abundance during the glacial interval. Above ~3.7 km there are
210 short maxima (a few cm) of *Cibicidoides* within or just above this interval, and
211 these are usually associated with maxima in *Uvigerina* and sometimes with *N.*
212 *umbonifera* maxima. However, there are no *Uvigerina* in the WEP at 4.25 km,
213 and there are no *N. umbonifera* above about 3.4 km water depth in the EEP. At
214 the two KNR 73 cores between 3.6 and 3.7 km in the EEP, *N. umbonifera* is
215 most abundant before 25 ^{14}C ka and in the Holocene.

216 The abundance of *G. ruber*, *G. sacculifer*, and total planktonic
217 foraminifera also show substantial variability (Figure 3). The two
218 *Globigerinoides* spp. represent about 10% of the planktonic fauna, on average,
219 and they reach a broad maximum in abundance during deglaciation.
220 Abundances are lower during the glacial and the Holocene (~upper 20 cm), and
221 (oddly) in the two cores with the thickest Holocene sections, the more solution-
222 susceptible *G. ruber* (Berger, 1970) is consistently least abundant.

223

224 3.2 Northwest Pacific.

225 Of the many ^{14}C dated sites in the NWP, results from two examples are
226 presented in Figure 4 (see also Cook and Keigwin, 2015). All cores from this
227 region contain abundant *Uvigerina* during deglaciation [Keigwin, 1998], but
228 abundance of this genus during the LGM is usually lower, and we have not
229 found clear glacial peaks in BF abundance deeper than 3.3 km on the northern
230 Emperor Seamounts. In both RNDB 11PC and Vinogradov 37 GGC (3225 and
231 3300 m, respectively, on Detroit Seamount) it was necessary to sample both the
232 working and the archived halves of the cores in order to obtain a datable
233 number of *Uvigerina*. Figure 4 shows only the 11 PC data, as an example. The
234 shape of the small LGM *Uvigerina* peak at ~110 cm is identical in each half of
235 the core, and $\delta^{18}\text{O}$ measurement of individual *Uvigerina*, the $\delta^{18}\text{O}$ on *N.*
236 *pachyderma s.*, and the dates on *N. pachyderma s.* all indicate that the small
237 LGM peak did not result from down core reworking of specimens from the
238 much larger deglacial peaks. The three deglacial peaks yield PF ^{14}C ages that
239 are all between 13.08 and 13.14 ^{14}C ka.

240 In contrast to the results at Detroit Seamount (50°N), Tenji Seamount
241 core RNDB 13PC (49°N) has two LGM BF peaks, based on $\delta^{18}\text{O}$ and PF ^{14}C
242 ages (Figure 4). This core also has three deglacial maxima in *Uvigerina*
243 abundance, with 4 PF ages of between 13.00 and 13.66 ^{14}C ka. Late in
244 deglaciation, 13PC has two additional BF abundance maxima, but the $\delta^{18}\text{O}$ data
245 do not reach the expected Holocene minimum of <3.5 ‰. The accompanying
246 trigger core (13PG) has only one late deglacial BF maximum and we consider
247 that core to be more reliable than the piston core because it captures the
248 expected Holocene $\delta^{18}\text{O}$ minimum.

249

250 4.0 DISCUSSION

251

252 Our observations of millennial scale changes in BF abundance in the EEP
253 are evidently new. The abundance patterns in Figure 2 have not been noted
254 previously, presumably because earlier sampling of EEP cores was done too
255 coarsely to resolve the millennial-scale events described here. (Most of the BF
256 peaks in the NWP and the EEP occur within a 5 cm interval of core.)
257 Furthermore, most previous authors have not measured the dry mass of their
258 sediment samples so it is impossible to know if there have been changes in
259 absolute BF abundance.

260 Despite the generally low sedimentation rates in our cores (all <7 cm/kyr
261 except for ME0005 24JC, >20 cm/kyr), where present, benthic foram
262 abundance maxima are several times larger than background values (Figure 2).
263 This suggests that they result from some process significant enough to survive
264 the effects of bioturbation. As bioturbation has often compromised detection of
265 possible millennial scale climate signals in the Equatorial Pacific, the strong BF
266 abundance signal documented here may hold promise for regional (and beyond)
267 correlation of events.

268 Below, we first consider possible contributions of surface ocean fertility
269 and sediment transport to observed BF abundance maxima, followed by a
270 discussion of the possible influences of dissolution and bioturbation on our ^{14}C
271 results. We then present a basin-wide compilation of glacial BF-PF ventilation
272 ages by depth that appears to indicate a widespread abyssal front in the glacial
273 Pacific.

274

275 4.1 Significance of BF abundance changes.

276

277 4.1.a. *BF abundance maxima and export production of organic carbon.*

278 A long history of studies concluded there must have been higher surface
279 ocean fertility in the EEP during glacial maxima (Arrhenius, 1952; Rea et al.,

280 1991; Herguera and Berger, 1991). At the eastern end of our study area, in the
281 Panama Basin, *Uvigerina* is thought to respond to increased export production
282 of carbon in terms of both increased size and abundance (Pedersen et al., 1988).
283 Loubere (1991) made a statistical regression of BF species percentages and
284 surface productivity in EEP core tops and found the fauna highly responsive to
285 productivity. In particular he noted that *Uvigerina*, *Cibicidoides* and *Melonis*
286 *barleeaanum* are associated with highest organic export fluxes close to the
287 Equator. Conversely, *N. umbonifera* was found to be typical of a lower
288 productivity assemblage that is more prominent south of 5°S. Based on
289 Loubere's (1991) work, our EEP results indicate highest productivity occurred
290 generally in the late glacial/early deglacial interval, a conclusion supported by
291 geochemical evidence for changing redox conditions (Berger et al., 1983).

292 In the Panama Basin, a broad deglacial maximum in the mass
293 accumulation rate of organic carbon and cooling of the sea surface has been
294 documented by Kienast et al. (2006), with a mid-point PF conventional ¹⁴C age
295 of ~13.9 ka. Far to the northwest on the Emperor Seamount chain, mass
296 accumulation rates of opal, calcium carbonate, organic carbon, and benthic
297 foraminiferal abundance, indicate higher fertility during the last deglaciation
298 compared to the LGM and the Holocene (Keigwin et al., 1992; Keigwin 1998).
299 High export production was confirmed by accumulation rates determined by Th
300 normalization in nearby cores (Crusius et al. 2004; Kohfeld and Chase, 2011),
301 and in the Gulf of Alaska by increased diatom export, Ba/Al, and CaCO₃
302 (Galbraith et al. 2007). Galbraith et al. (2007) argued that these changes
303 occurred abruptly in the Gulf of Alaska at the onset of the Bolling/Allerod
304 interval in the North Atlantic region (~14.5 cal. ka), with ¹⁴C age constraints
305 from loosely bracketing conventional PF ages of 13.0 and 15.1 ¹⁴C ka.

306 Among our EEP cores, PLDS 7G, VNTR01 10GC and KNR73-3PC
307 contain BF abundance maxima with PF ¹⁴C ages of ~14.15, 14.18 and 14.73
308 ¹⁴C ka, respectively (Fig. 2) that indicate that they may be broadly synchronous
309 with the previously documented productivity increase in the Panama Basin
310 (Kienast et al., 2006), after allowing for possible regional gradients of deglacial
311 reservoir age of several hundreds of years (Lindsay et al. 2015.; Rae et al.
312 2014). Similarly, the multiple BF abundance peaks with PF ¹⁴C ages of 13.0 to
313 13.6 ¹⁴C ka in our NWP cores (Fig. 4) may be related to the sudden increase in
314 productivity in the northern Pacific (Keigwin et al. 1992; Galbraith et al. 2007;
315 Kohfeld and Chase, 2011).

316 In the four EEP cores in which we have counted planktonic foram
317 abundance, highest abundances generally occur during deglaciation, but the
318 broad deglacial rise in planktonic species does not usually contain abundant
319 benthic taxa (Fig. 3). Although planktonic foraminifera are most abundant

320 during the deglacial interval, we cannot be sure to what extent this reflects
321 preservation rather than productivity (Berger, 1992, for example), or sediment
322 focusing (Marcantonio et al., 2001; Kienast et al., 2006).

323

324 4.1.b. *Sediment transport*. Although there is strong evidence for high
325 productivity in the NWP during deglaciation and in the EEP during glaciation,
326 it remains possible that peaks in BF abundance could be an artifact of extensive
327 winnowing of clay and silt. In the case of extreme and persistent winnowing,
328 BF peaks could be created. However, in order to explain increases in benthic
329 abundance by factors of 5 or 10, such as during the Holocene or the events at
330 ~ 14.7 and 15.5 ^{14}C ka in cores 3PC and 4PC, respectively (Figure 2),
331 enrichment by winnowing would have necessarily produced substantially
332 condensed intervals. Our dating of these cores was not designed to test this
333 possibility, but the depth distribution of PF ^{14}C ages in 3PC and 4PC is
334 inconsistent with the presence of large discontinuities. At the Emperor
335 Seamounts, each site has multiple deglacial peaks in *Uvigerina* associated with
336 similar PF ^{14}C ages of ~ 13.1 ^{14}C ka, but the maximum abundances at core 11PC
337 are $\sim 50/\text{g}$ whereas at core 13PC it is $\sim 28/\text{g}$. If these events resulted from
338 winnowing, then one would expect more winnowing at 11PC, yet based on the
339 *Uvigerina* $\delta^{18}\text{O}$, the two cores have similar rates of sedimentation.

340

341 4.2. Planktonic age bias due to selective dissolution and bioturbation.

342

343 Barker et al. (2007) noted that the effects of selective dissolution of
344 surface dwelling planktonic species such as *G. sacculifer* and *G. ruber*
345 compared to deeper-dwelling planktonics, variable residence time of shells in
346 the sediment surface mixed layer, and low rates of sedimentation can combine
347 to create ^{14}C age offsets between different PF species from the same sediment
348 sample. The conceptual model predicts that relatively dissolution prone species
349 will show a young ^{14}C age bias with respect to dissolution resistant species, as a
350 result of selective removal of older tests from the mixed layer population.
351 Broecker and Clark (2011) proposed that this effect could lead to artificially
352 large ventilation ages that are based on the difference between BF ^{14}C dates and
353 PF ^{14}C dates from co-deposited dissolution prone surface dwelling species. For
354 example, they noted that in four of seven deep water core tops from the
355 equatorial Pacific that are today bathed by calcite undersaturated water, *G.*
356 *sacculifer* has a ^{14}C age about 1000 years younger than solution-resistant
357 planktonic species. Comparison of PF ages from paired measurements of *G.*
358 *ruber* and *G. sacculifer* in cores 3PC and 4PC provide some insight into the
359 importance of this process in the deep EEP as, even though both species are

360 surface-dwellers, *G. ruber* is known to be less resistant to dissolution than *G.*
361 *sacculifer* (Berger, 1970).

362 We find no statistically meaningful inter-species age difference across all
363 available *G. ruber* and *G. sacculifer* measurement pairs ($n = 17$, Table 3). If we
364 break the pairs out according to time intervals (based on average PF ^{14}C ages)
365 corresponding broadly to the Holocene (0-10 ^{14}C ka), deglaciation (10-15 ^{14}C
366 ka), and the glacial period (15-25 ^{14}C ka), we find that only for the deglacial
367 period is *G. ruber* measurably younger than *G. sacculifer* (-233 ± 112 ^{14}C yr, n
368 $= 6$). For the glacial interval, the relationship is marked by large variation in
369 both magnitude and sign ($+59 \pm 745$ ^{14}C yr, $n = 7$). These results, along with an
370 analysis of the absolute deviation of the inter-species differences (Table 3),
371 suggest that the observed age differences are largely randomized by
372 bioturbation and generally do not preserve systematic offsets that might arise
373 from differential dissolution in the mixed layer or from ecological habitat
374 preferences (which are expected to be similar in the EEP for these two species,
375 for example, Watkins et al. 1996). The results are, furthermore, largely
376 insensitive to the placement of age boundaries in the analysis.

377

378 We also note that the effect described by Barker et al. (2007) is based on
379 modern ocean observations and late Holocene sediment that reflect especially
380 corrosive conditions on the sea floor. However, glacial and deglacial sediment
381 is generally thought to be better preserved than Holocene sediment in the
382 Equatorial Pacific. All other things being equal, the impacts of selective
383 dissolution and young PF age biases should then be greater in the Holocene
384 than during the glaciation and deglaciation, reducing rather than enhancing
385 observed differences in ventilation age that might otherwise arise due to relative
386 aging of Pacific deep water during glacial times. Our Holocene observations
387 are limited to only two cores (3, 4PC), but with BF-PF ventilation ages of 1000
388 to 2000 years, they are close to the expected values based on modern
389 observations (~ 1600 yrs; Broecker et al., 1984). As we describe below, BF-PF
390 ventilation ages in the glacial and early deglacial deep Equatorial Pacific are
391 much higher and more variable. In our view, the largest contribution to
392 uncertainty is from bioturbation which, for PFs, we estimate conservatively to
393 be on the order of ± 500 yr based on absolute deviations of PF age pairs for the
394 glacial period (Table 3). Since we have sampled BFs at local peaks of absolute
395 abundance, uncertainties in BF ages from sediment mixing should be
396 minimized, and are likely more than adequately captured by differences in
397 ventilation age for different BF species sampled at or near the same
398 stratigraphic level (Fig. 2, Fig. 5).

399

400

401 4.3. Equatorial and North Pacific BF-PF ventilation age.

402 There is general agreement that there were significant changes in the
403 deep circulation of the North Pacific during the past ~25 ka, as evident in
404 various proxy data (Boyle and Keigwin, 1985; Zahn et al., 2001; Keigwin,
405 1998; Ohkushi et al., 2003; Galbraith et al., 2007; Okazaki et al. 2010; Lund et
406 al. 2011). Of the many proxies brought to bear on the subject, ^{14}C has been of
407 recent interest because of the suggestion that the deep ocean may have driven
408 the deglacial atmospheric pCO_2 increase and $\Delta^{14}\text{C}$ decrease if long-sequestered
409 CO_2 was released to the atmosphere as deep ocean ventilation increased after
410 the glacial period (Hughen et al., 2004; Broecker et al. 2004; Marchitto et al.
411 2007; Skinner et al. 2010). Okazaki et al. (2010) compiled available BF -PF ^{14}C
412 age pairs from North Pacific locations and concluded that BF-PF ventilation
413 ages began to decrease beginning 20-19 calendar ka and remained ~500-1000
414 ^{14}C yr lower than during the early LGM until the end of Heinrich Event 1 (H-1,
415 17.5 to 15 cal. ka), suggesting that deep water may have been produced
416 somewhere in the region to depths as great as 3 km during the interval. More
417 recently, Rae et al. (2014) showed that ventilation of the deep NEP (3640 m)
418 improved suddenly during “the middle of H-1” based on a brief but seemingly
419 robust collapse of the age difference in two BF-PF pairs. ^{14}C results from
420 another NEP core indicate that ventilation at 2.7 km was the same during the
421 last glaciation as today (Lund et al. 2011). On the other hand, Jaccard and
422 Galbraith (2013) suggest that bottom waters at 2400 m (and deeper) were
423 poorly ventilated until ~ 15 calendar ka based on measurements of authigenic
424 uranium in NWP core 13PC. Aside from four BF-PF pairs of Shackleton et al.
425 (1988) and the previous work of and Broecker et al. (2004, 2008) and Broecker
426 and Clark (2010), ours are the only other deep water ventilation estimates from
427 the Equatorial Pacific.

428 Whereas Okazaki et al. (2010) compiled ventilation data from the North
429 Pacific in a time series, here we compile the data they used along with more
430 recently published data and our new results into a single depth reconstruction,
431 but with distinct labels for results from the NWP, NEP, EEP and WEP (Figure
432 5). In our analysis we consider primarily the BF-PF ventilation age, since this
433 metric does not involve assumptions regarding reservoir age effects or
434 calibration to calendrical age, as may be needed to estimate physical ventilation
435 age using decay trajectories (c.f. Adkins and Boyle, 1997). Thus, plotted results
436 are not subject to potentially significant uncertainties regarding the magnitude
437 and spatial distribution of past surface reservoir ages, but ventilation ages with
438 respect to the coeval atmosphere may be more variable than indicated by the
439 reconstruction based on BF-PF age.

440 We also lump together all results for which any one of the associated
441 conventional PF ^{14}C ages lies within 1σ of 15 to 25 ^{14}C ka, a relatively broad
442 interval that includes both the LGM and H-1. The approach is deliberately
443 inclusive so as to maximize the number of (still relatively sparse) observations
444 under consideration and because climatically and oceanographically significant
445 events documented elsewhere are not always well resolved in the Pacific
446 Ocean. For example H-1 is not directly evident in the Pacific as an armada of
447 icebergs and is otherwise difficult to identify with certainty. Furthermore, as has
448 been noted previously (Keigwin et al., 1992; Skinner and Shackleton, 2005),
449 Pacific BF $\delta^{18}\text{O}$ data lag Atlantic data because of the long interbasin transit
450 time, so from that perspective the end of the LGM in the deep Pacific is
451 younger than in the Atlantic. Glaciological evidence from Hawaii indicates the
452 LGM persisted in the North Pacific until about 15 ka (Blard et al., 2007).

453 The glacial BF-PF ventilation age results are given in Figure 5 along with
454 an estimate of the basin-wide average $\Delta^{14}\text{C}$ in modern seawater (solid red line,
455 upper X axis scale) by depth from the World Ocean Circulation Experiment
456 (<http://cdiac.ornl.gov/oceans/glodap/>), and were averaged over the North
457 Pacific between the Equator and 55°N by R. M. Key (pers. comm.). Data used
458 in making Figure 5 are listed as SOM Table 1. Equivalent sea water ^{14}C ages
459 (open red symbols) are effectively relative to that of the pre-nuclear, pre-
460 industrial atmosphere and have been corrected by 400 yrs for presentation on
461 the BF-PF age scale (lower X axis). Formal uncertainties for the apparent
462 ventilation ages were described earlier and are listed in Table 2; a more realistic
463 assessment may be provided by the observed spread in BF-PF age differences
464 for different cores from similar depths in the same ocean region or, (for the new
465 results) for different BF species sampled at or near the same stratigraphic level.
466 Since the glacial data span a relatively large age interval, some of the observed
467 spread may reflect authentic time dependent variability of BF-PF ventilation
468 age. As we will show below, however, plausible time-dependent variations (i.e.,
469 Okazaki et al. 2010) do not appear to have influenced the overall structure of
470 the BF-PF depth profile.

471 Despite the observed scatter, the glacial age sediment data are roughly
472 consistent with the modern sea water ^{14}C profile at depths above ~ 3.2 km
473 (Figure 5). A distinctive break relative to the modern profile occurs about 3.1
474 km in the sediment results from the Equatorial Pacific; data between 1.9 and 3
475 km average 1610 ± 250 yrs, whereas deeper data fall consistently on the “old”
476 side of the modern profile, with an average BF-PF ventilation age of
477 3080 ± 1125 yrs. Taken at face value, these results record a pronounced abyssal
478 front between shallower, well ventilated waters and deeper, poorly ventilated
479 waters in the Equatorial Pacific at a modern water depth of 3.0 to 3.2 km. The

480 impression of a front is reinforced by consistency of equatorial Pacific and
481 North Pacific results above 3.0 km. Four BF-PF pairs from 3.2 km in the EEP
482 (Shackleton et al. 1988) are consistent with a front at or just above this depth
483 (SOM Fig. 2), but are excluded from Fig. 5 since, unlike the other EEP data,
484 they are based on PF measurements in *N. dutertrei* which may be biased old
485 (potentially reducing BF-PF with respect to other data shown) (Barker et al.,
486 2007). Inclusion of these data would yield a deep Equatorial Pacific BF-PF
487 average of 2890 ± 1100 yr.

488 In order to evaluate the possible influence of time-dependent variation in
489 our reconstruction, we separate glacial results into two age classes, greater or
490 less than $\sim 19.6 \pm 0.4$ calendar ka, corresponding to the transition to younger
491 BF-PF ventilation age documented above ~ 3 km in the North Pacific (Okazaki
492 et al. 2010). This age break corresponds to conventional PF ^{14}C ages greater or
493 less than 17.0 ^{14}C ka in our results (Table 2), with an age spread after
494 calibration that incorporates large uncertainty in the surface reservoir age (ΔR
495 of 0-600 in CALIB 7.1, after Stuiver and Reimer 1993) but which nevertheless
496 securely identifies samples older than the transition indicated by Okazaki et al.
497 (2010). For the deep Equatorial Pacific, there is no statistically significant
498 difference of BF-PF age for pre- and post transition samples; “pre-transition”
499 samples are 2850 ± 870 yr (n=9) and “post-transition” samples are 3440 ± 1440
500 yr (n=6) and 2920 ± 750 yr (n=5) when the oldest (6050 yr) sample is excluded.
501 Most importantly, the large contrast with results above and below 3.1 km is
502 maintained both within and across the two age classes, as can be seen in
503 separately annotated symbols in Figure 5. We also note that some of the time-
504 dependent variation reconstructed by Okazaki et al. (2010) may relate to
505 changes in the range of depths represented by their core selection, since the
506 largest BF-PF ages in the North Pacific appear to be in “pre-transition” samples
507 from 2.4 to 2.9 km.

508 Any possible northern manifestation of poorly ventilated waters within
509 the deep N. Pacific remains difficult to evaluate from our compilation of results.
510 Above ~ 3.1 km, both the northern and Equatorial Pacific measurements indicate
511 similar BF-PF ventilation ages. This is likely a robust finding, since similar
512 methods were employed in both areas (most NWP results used our methods;
513 Cook and Keigwin, 2015). Below ~ 3.1 km, there is a paucity of glacial BF-PF
514 age pairs from the northern N. Pacific that meet our criteria for co-deposition
515 (within 1 cm). Of these, 2 come from the relatively brief “middle H-1”
516 ventilation event documented in the NEP by Rae et. al. (2014) and, they
517 suggest, these were associated with a dramatic increase in the surface ocean
518 reservoir age (leading to under-representation of the actual deep water ^{14}C age
519 from BF-PF age pairs). We note that many of the deepest northern N. Pacific

520 cores come from areas where today the surface ocean reservoir age appears to
521 be locally elevated (see map of core depths vs. estimated pre-nuclear surface
522 water ^{14}C age in Supplementary online Figure 1. If the estimated recent, pre-
523 nuclear reservoir age distribution (Rubin and Key, 2002) reflects the potential
524 for locally- or regionally- anomalous increases in reservoir age during the
525 glacial and/or deglacial periods, results from deeper N. Pacific cores in Fig. 5
526 may under-represent actual deep water ^{14}C ages with respect to the larger
527 number of records available from above ~ 3.1 km. Nonetheless, there is as yet
528 no clear ^{14}C evidence for excessive glacial aging of deep waters in the northern
529 N. Pacific. The single new BF-PF age pair characterizing the deep NWP, as
530 noted in section 3.2, ultimately required sampling of the archive half of
531 Vinogradov 37GGC in order to obtain enough *Uvigerina* for dating. Given that
532 this is the only result for the deep NWP, more work in the region is clearly
533 needed.

534 Radiocarbon data from the LGM of the subtropical western North
535 Atlantic indicate apparent ventilation ages of only about 1500 years in cores as
536 deep as 4.7 km (Keigwin, 2004). Evidently the very old deep water we find in
537 the Equatorial Pacific Ocean did not extend substantially unmixed into the deep
538 western North Atlantic. In the South Atlantic at ~ 3.8 km Skinner et al. (2010)
539 also measured LGM BF-PF ventilation ages of about 1500 years. However,
540 they also identified large changes in the surface reservoir age, suggesting that
541 that deep South Atlantic benthic ^{14}C ages were 3500 to 4000 yrs older than the
542 coeval atmosphere during the LGM. Values in this range were maintained for
543 ~ 6 kyr, with a maximum at ~ 19 calendar ka BP. In Figure 5 we show prior
544 estimates of $\Delta^{14}\text{C}$ (vs. coeval atmosphere) for these data as a box centered at 3.8
545 km depth to represent a likely end-member value for Lower Circumpolar Deep
546 Water (LCDW) (Burke and Robinson, 2012). Projection of these results from
547 the $\Delta^{14}\text{C}$ (vs. atmosphere) axis onto the BF-PF age axis carries some
548 uncertainties related to possible local deviations from mean ocean surface
549 reservoir age, but the overall impression is that both the deep Equatorial Pacific
550 and deep Southern Ocean were comparably aged during much of the LGM. We
551 thus posit a direct connection between the two locations via mixing with
552 LCDW as in the present day ocean, implying that similarly old waters may have
553 had a circumpolar distribution during part or all of the glacial period.

554 Thus, we argue that both the Southern Ocean (Skinner et al. 2015), and
555 much of the world's most voluminous ocean basin harbored a potentially vast
556 reservoir of old carbon. Broecker et al. (2004) estimated that the volume and
557 degree of deep ocean ^{14}C depletion needed to account for the equilibrium
558 change in $\Delta^{14}\text{C}$ of glacial atmosphere and remaining portions of the ocean was
559 much larger than seemed plausible given existing observational constraints.

560 However, Burke and Robinson (2012) suggest that recent observations of
561 abyssal depletion during the LGM, comparable in degree to those documented
562 here, would be sufficient to account for subsequent transient depletion of the
563 atmosphere and surface ocean during deglaciation. The complementary
564 observations of deglacial atmospheric $\Delta^{14}\text{C}$ and CO_2 change (Machitto et al.
565 2007) suggest a common cause relating to improved ventilation of the ocean's
566 deepest waters (Toggweiler et al., 2006).

567

568

569 CONCLUSIONS

570 Our study of deep ocean ventilation ages based on the difference in ^{14}C
571 ages of coexisting benthic and planktonic foraminifera in cores from the deep
572 Equatorial Pacific and Northwest Pacific reveals the following:

573

574 1. Maxima in absolute abundance of benthic foraminifera are common in both
575 the NWP and most Equatorial Pacific sites during the Holocene, the
576 deglaciation, and the glacial period. They appear to be related to higher fertility
577 during glaciation in the EEP and during deglaciation in both regions.

578 Abundance of planktonic foraminifera is lowest in the late Holocene,
579 intermediate during the glacial, and highest during deglaciation.

580

581 2. In the EEP, paired ^{14}C measurements of *G. sacculifera* and *G. ruber* do not
582 reveal systematic differences of age that might arise from ecological
583 preferences or differential dissolution in the sediment mixed layer. Variations
584 in the sign and magnitude of inter-species age differences can be attributed
585 largely to bioturbation.

586

587 3. Our few Holocene apparent ventilation ages from the EEP fall in the 1000 to
588 2000 year range, similar to today. However, deglacial and glacial values are
589 both larger and more variable (1500 to 6000 years), with a mean of ~ 3100 yrs
590 that is greater than anything observed previously in the shallower Equatorial
591 and North Pacific. Data above about 2 km indicate slightly younger ventilation
592 ages compared to the modern (GLODAP) mean sea water depth ^{14}C profile for
593 the same region. A large contrast of apparent ventilation age (~ 1500 yrs) above
594 and below 3.1 km in the Equatorial Pacific is emphasized by comparison to the
595 modern data, suggesting a front in BF-PF ventilation age near that depth.

596

597 4. The large BF-PF ^{14}C differences for >3.1 km in the glacial equatorial
598 Pacific indicate bottom waters that may be comparably aged to the Lower
599 Circumpolar Deep Water in the South Atlantic. Thus, there may have been a

600 circumpolar distribution of old water that extended as far as the equatorial Pacific
601 and that was likely sufficient to have caused the transient decrease in
602 atmospheric radiocarbon activity during deglaciation.

603

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REFERENCES

621

- 622 Adkins, J. F. and E. A. Boyle (1997). "Changing atmospheric $\Delta^{14}\text{C}$ and the record of deep
623 water paleoventilation ages." Paleoceanography **12**: 337-344.
- 624 Anderson, R., S. Ali, et al. (2009). "Wind-driven upwelling in the Southern Ocean and the
625 deglacial rise in atmospheric CO_2 ." Science **323**: 1443-1447.
- 626 Arrhenius, G. (1952). "Sediment cores from the east Pacific." Rep. Swedish Deep Sea
627 Exped. **5**: 1-228.
- 628 Barker, S., W. Broecker, et al. (2007). "Radiocarbon age offsets of foraminifera resulting
629 from differential dissolution and fragmentation within the sedimentary
630 bioturbated zone." Paleoceanography **22**: PA2205.
- 631 Barnola, J. M., D. Raynaud, et al. (1987). "Vostok ice core provides 160,000-year record
632 of atmospheric CO_2 ." Nature **329**: 408414.
- 633 Berger, W. (1970). "Planktonic foraminifera: selective solution and the lysocline."
634 Marine Geol. **8**: 111-138.
- 635 Berger, W. (1992). Pacific carbonate cycles revisited: arguments for and against
636 productivity control. Centenary of Japanese Micropaleontology. K. Ishizaki and
637 T. Saito. Tokyo, Terra Scientific Publishing Co.: 15-25.
- 638 Berger, W. H., R. C. Finkel, et al. (1983). "Glacial-Holocene transition in deep sea
639 sediments: Manganese-spike in the east equatorial Pacific." Nature **303**: 231-
640 233.

641 Blard, P.-H., J. Lave, et al. (2007). "Persistence of full glacial conditions in the central
642 Pacific until 15,000 years ago." Nature **449**: 591-595.

643 Boyle, E. A. (1988). "Vertical oceanic nutrient fractionation and glacial/interglacial CO₂
644 cycles." Nature **331**: 55-56.

645 Boyle, E. A. and L. D. Keigwin (1985/86). "Comparison of Atlantic and Pacific
646 paleochemical records for the last 215,000 years: changes in deep ocean
647 circulation and chemical inventories." Earth and Planet. Sci. Lett. **76**: 135-150.

648 Broecker, W. and S. Barker (2007). "A 190 ‰ drop in atmosphere's D14C during the
649 "Mystery Interval" (17.5 to 14.5 kyr)." Earth Planet. Sci. Lett. **256**: 90-99.

650 Broecker, W., S. Barker, et al. (2004). "Ventilation of the glacial deep Pacific Ocean."
651 Science **306**: 1169-1172.

652 Broecker, W. and E. Clark (2010). "Search for a glacial-age 14C-depleted ocean
653 reservoir." Geophys. Res. Lett. **37**.

654 Broecker, W. and E. Clark (2011). "Radiocarbon-age differences among coexisting
655 planktic foraminifera shells: The Barker Effect." Paleoceanography **26**: PA2222.

656 Broecker, W., E. Clark, et al. (2008). "Near constancy of the Pacific Ocean surface to
657 mid-depth radiocarbon-age difference over the last 20 kyr." Earth Planet. Sci.
658 Lett. **274**: 322-326.

659 Broecker, W. S., A. Mix, et al. (1984). "Radiocarbon measurements on coexisting benthic
660 and planktic foraminifera shells: potential for reconstructing ocean ventilation
661 times over the past 20,000 years." Nuclear Inst. and Methods in Phys. Res.
662 section B: Beam Interactions with Materials and Atoms **5**: 332-339.

663 Bryan, S. P., T. M. Marchitto, et al. (2010). "The release of 14C-depleted carbon from the
664 deep ocean during the last deglaciation: Evidence from the Arabian Sea." Earth
665 Planet. Sci. Lett. **298**: 244-254.

666 Burke, A. and L. F. Robinson (2012). "The Southern Ocean's role in carbon exchange
667 during the last deglaciation." Science **335**: 557-561.

668 Cleroux, C., P. deMenocal, et al. (2011). "Deglacial radiocarbon history of tropical
669 Atlantic thermocline waters: absence of CO₂ reservoir purging signal." Quat. Sci.
670 Rev. **30**: 1875-1882.

671 Cook, M. and L. Keigwin (2015). "Radiocarbon profiles of the NW Pacific from the LGM
672 and deglaciation: Evaluating ventilation metrics and the effect of uncertain
673 surface reservoir ages." Paleoceanography **30**, doi:10.1002/2014PA002649.

674 Crusius, J., T. F. Pedersen, et al. (2004). "Elevated productivity in the NW Pacific during
675 the Bolling warm period as a cause of reduced intermediate-water oxygen
676 concentrations along the California margin." Geology **32**: 633-636.

677 Davies-Walczak, M., A. C. Mix, et al. (2014). "Late glacial to Holocene radiocarbon
678 constraints on North Pacific Intermediate Water ventilation and deglacial
679 atmospheric CO₂ sources." Earth and Planetary Science Letters **397**: 57-66.

680 dePol-Holz, R., L. Keigwin, et al. (2010). "No signature of abyssal carbon in intermediate
681 waters off Chile during deglaciation." Nature Geosciences **3**: 192-195.

682 Duplessy, J. C., N. J. Shackleton, et al. (1988). "Deepwater source variations during the
683 last climatic cycle and their impact on the global deepwater circulation."
684 Paleoceanography **3**: 343-360.

685 Galbraith, E. D., S. L. Jaccard, et al. (2007). "Carbon dioxide release from the North
686 Pacific abyss during the last deglaciation." Nature **449**: 890-894.

687 Herguera, J. C. and W. H. Berger (1991). "Paleoproductivity from benthic foraminifera
688 abundance: Glacial to postglacial change in the west-equatorial Pacific." Geology
689 **19**: 1173-1176.

690 Hughen, K., S. Lehman, et al. (2004). "14C Activity and global carbon cycle changes over
691 the past 50,000 years." Science **303**: 202-207.

692 Jaccard, S. and E. Galbraith (2013). "Direct ventilation of the North Pacific did not reach
693 the deep ocean during the last deglaciation." Geophys. Res. Lett. **40**:
694 doi:10.1029/2012GL054118.

695 Keigwin, L. D. (1998). "Glacial-age hydrography of the far northwest Pacific Ocean."
696 Paleoceanography **13**: 323-339.

697 Keigwin, L. D. (2004). "Radiocarbon and stable isotope constraints on last glacial
698 maximum and Younger Dryas ventilation in the western North Atlantic."
699 Paleoceanography **19**: doi: 10.1029/2004PA001029.

700 Keigwin, L. D. and G. A. Jones (1990). "Deglacial climatic oscillations in the Gulf of
701 California." Paleoceanography **5**: 1009-1023.

702 Keigwin, L. D., G. A. Jones, et al. (1992). "A 15,000 year paleoenvironmental record from
703 Meiji Seamount, far northwestern Pacific." Earth Planet. Sci. Lett. **111**: 425-440.

704 Kienast, M., S. S. Kienast, et al. (2006). "Eastern Pacific cooling and Atlantic overturning
705 circulation during the last deglaciation." Nature **443**: 846-849.

706 Kohfeld, K. E. and Z. Chase (2011). "Controls on deglacial changes in biogenic fluxes in
707 the North Pacific Ocean." Quat. Sci. Rev. **30**: 3350-3363.

708 Lindsay, C., S. Lehman, et al. (2015). "The surface expression of radiocarbon anomalies
709 near Baja California during deglaciation." Earth Planet. Sci. Lett. **422:67-74**.

710 Loubere, P. (1991). "Deep-sea benthic foraminiferal assemblage response to a surface
711 ocean productivity gradient: A test." Paleoceanography **6**: 193-204.

712 Lund, D. C., A. C. Mix, et al. (2011). "Increased ventilation age of the deep northeast
713 Pacific Ocean during the last deglaciation." Nature Geoscience **4**: 771-774.

714 Marcantonio, F., R. F. Anderson, et al. (2001). "Sediment focusing in the central
715 Equatorial Pacific Ocean." Paleoceanography **16**: 260-267.

716 Marchitto, T. M., S. J. Lehman, et al. (2007). "Marine radiocarbon evidence for the
717 mechanism of deglacial atmospheric CO₂ rise." Science **316**: 1456-1459.

718 Ohkushi, K., T. Itaka, et al. (2003). "Last glacial-Holocene change in intermediate water
719 ventilation in the Northwestern Pacific." Quat. Sci. Rev. **22**: 1477-1484.

720 Okazaki, Y., A. Timmermann, et al. (2010). "Deepwater formation in the North Pacific
721 during the last glacial termination." Science **329**: 200-204.

722 Pedersen, T. F., M. Pickering, et al. (1988). "The response of benthic foraminifera to
723 productivity cycles in the eastern Equatorial Pacific: Faunal and geochemical
724 constraints on glacial bottom water oxygen levels." Paleoceanography **3**: 157-
725 168.

726 Petit, J. R., J. Jouzel, et al. (1999). "Climate and atmospheric history of the past 420,000
727 years from the Vostok ice core, Antarctica." Nature **399**: 429-436.

728 Rae, J. W. B., M. Sarnthein, et al. (2014). "Deep water formation in the North Pacific and
729 deglacial CO₂ rise." Paleoceanography **29(6)**: 645-667.

730 Rea, D. K., N. G. Pisias, et al. (1991). "Late Pleistocene paleoclimatology of the central
731 Equatorial Pacific: Flux patterns of biogenic sediments." Paleoceanography **6**:
732 227-244.

733 Robinson, L. F., J. F. Adkins, et al. (2005). "Radiocarbon variability in the western North
734 Atlantic during the last deglaciation." Science **310**: 1469-1473.

735 Rubin, S. and R. Key (2002). "Separating natural and bomb-produced radiocarbon in the
736 ocean: The potential alkalinity method." Global Biogeochemical Cycles **16**: 1105,
737 doi:1110.1029/2001GB001432.

738 Shackleton, N. J., J. C. Duplessy, et al. (1988). "Radiocarbon age of last glacial Pacific deep
739 water." Nature **335**: 708-711.

740 Sigman, D. M., M. P. Hain, et al. (2010). "The polar ocean and glacial cycles in
741 atmospheric CO₂ concentration." Nature **466**(7302): 47-55.

742 Sikes, E. L., C. R. Samson, et al. (2000). "Old radiocarbon ages in the southwest Pacific
743 Ocean during last glacial period and deglaciation." Nature **405**.

744 Skinner, L., I. N. McCave, et al. (2015). "Reduced ventilation and enhanced magnitude of
745 the deep Pacific carbon pool during the last glacial period." Earth Planet. Sci. Lett.
746 **411**: 45-52.

747 Skinner, L. C., S. Fallon, et al. (2010). "Ventilation of the deep Southern Ocean and
748 deglacial CO₂ rise." Science **328**(5982): 1147-1151.

749 Skinner, L. C. and N. J. Shackleton (2005). "An Atlantic lead over Pacific deep-water
750 change across Termination 1: implications for the application of the marine
751 isotope stage stratigraphy." Quat. Sci. Rev. **24**: 571-580.

752 Stott, L., J. Southon, et al. (2009). "Radiocarbon age anomaly at intermediate water
753 depth in the Pacific Ocean during the last deglaciation." Paleoceanography **24**:
754 PA2223.

755 Stuiver, M. and H. Polach (1977). "Reporting of ¹⁴C data." Radiocarbon: 355-363.

756 Stuiver, M. and P. J. Reimer (1993). "Extended ¹⁴C database and revised CALIB
757 radiocarbon calibration program." Radiocarbon **35**: 215-230.

758 Toggweiler, J. R., J. L. Russell, et al. (2006). "Midlatitude westerlies, atmospheric CO₂,
759 and climate change during the ice ages." Paleoceanography **21**(2).

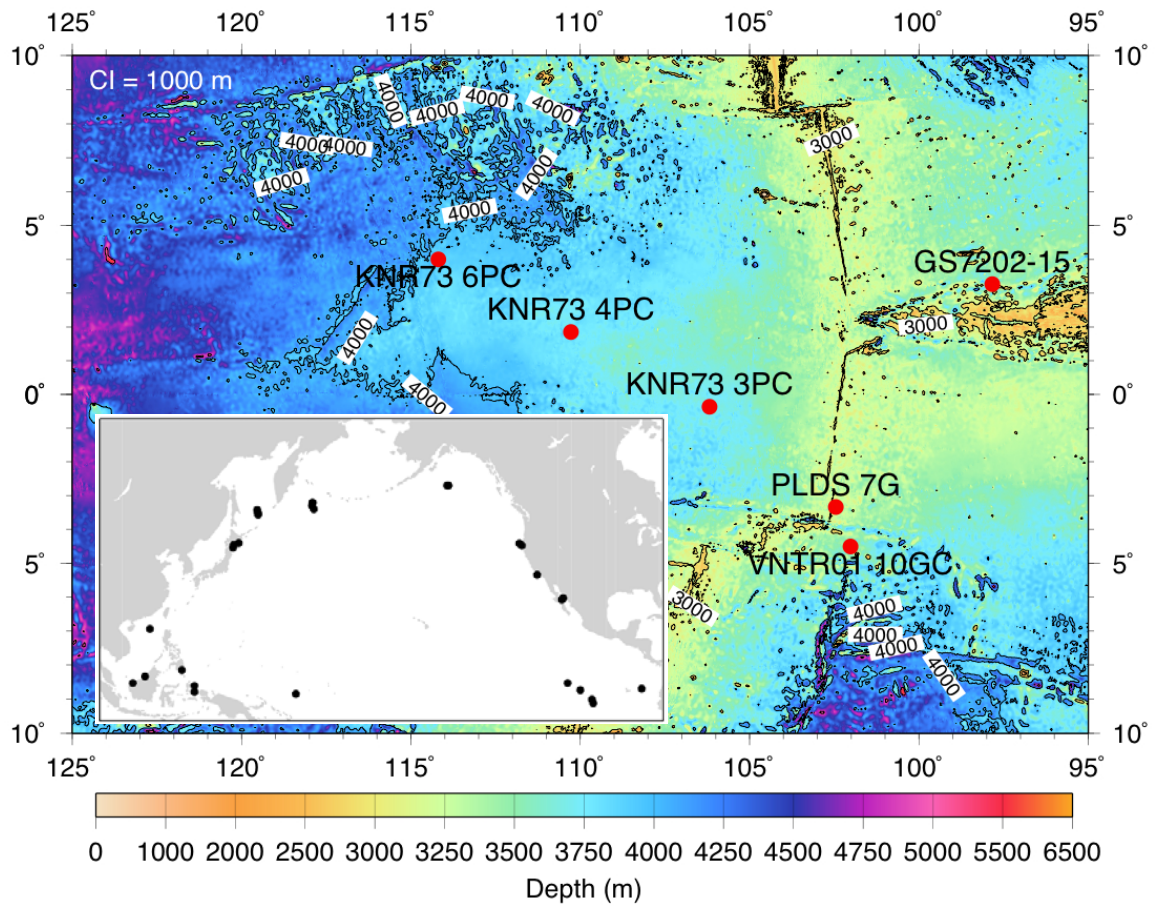
760 Watkins, J. M., A. C. Mix, et al. (1996). "Living planktic foraminifera: tracers of circulation
761 and productivity regimes in the central equatorial Pacific." Deep-Sea Res. **43**:
762 1257-1282.

763 Zahn, R., T. F. Pedersen, et al. (1991). "Water mass conversion in the glacial Subarctic
764 Pacific (54N, 148W): Physical constraints and the benthic-planktonic stable
765 isotope record." Paleoceanography **6**: 543-560.

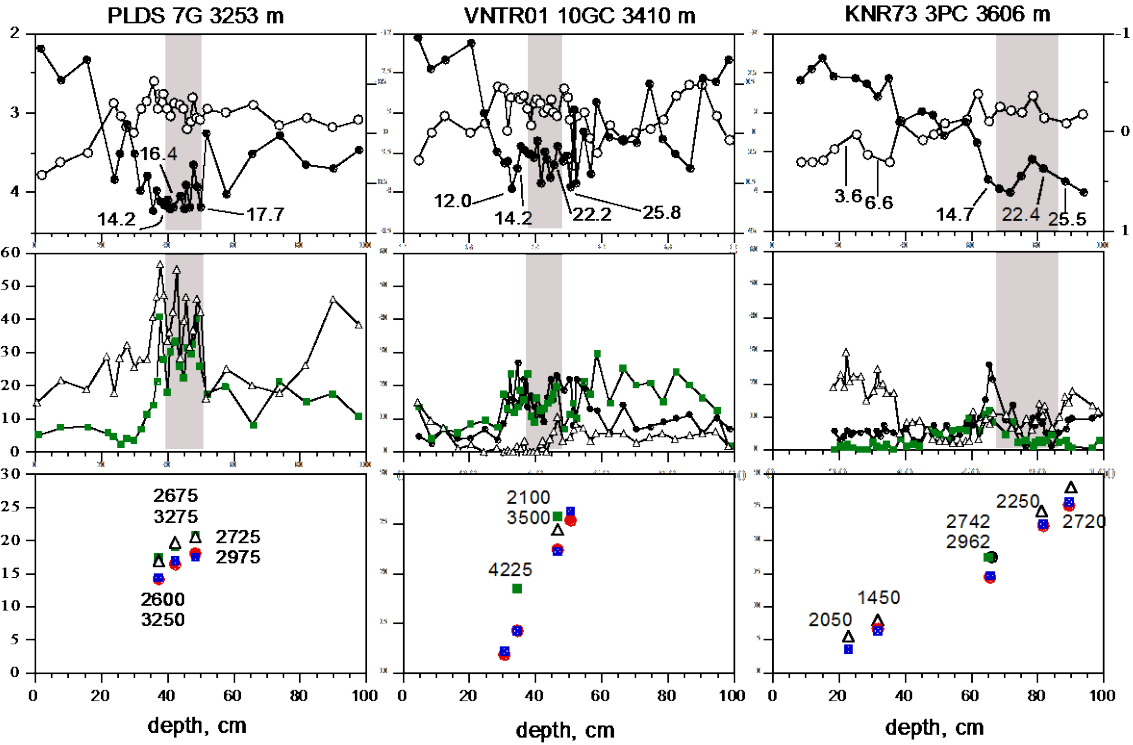
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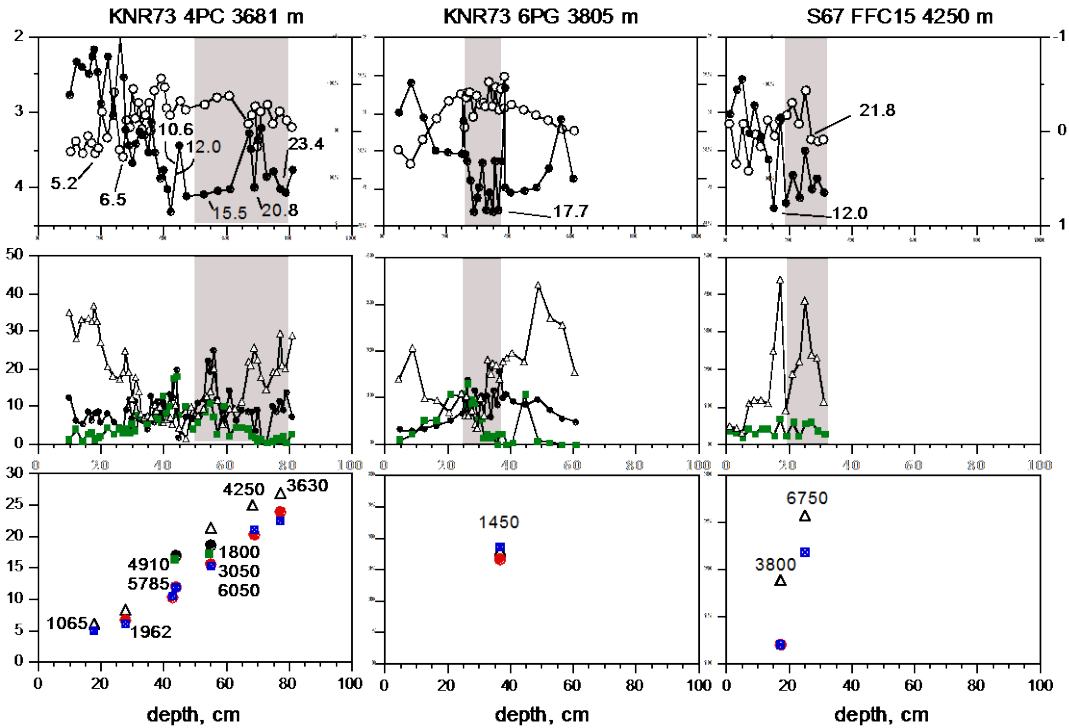
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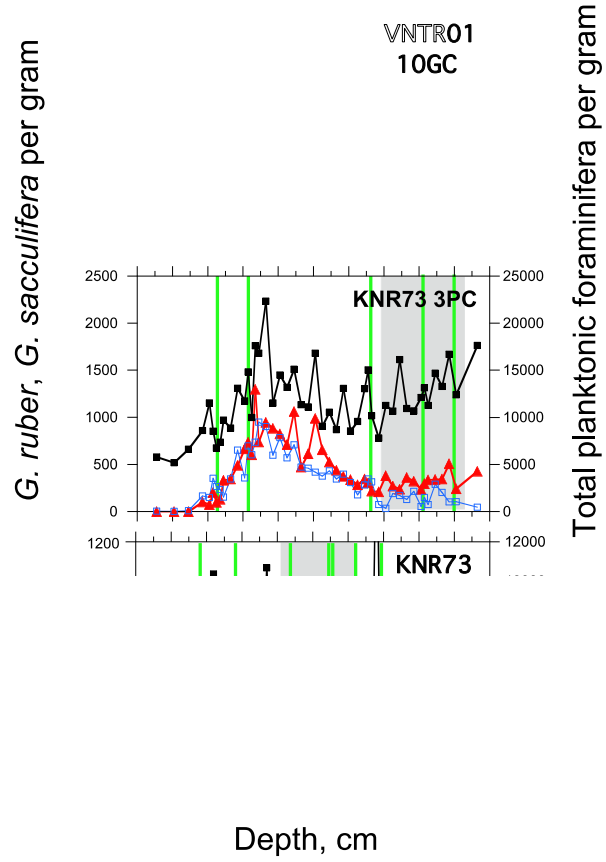
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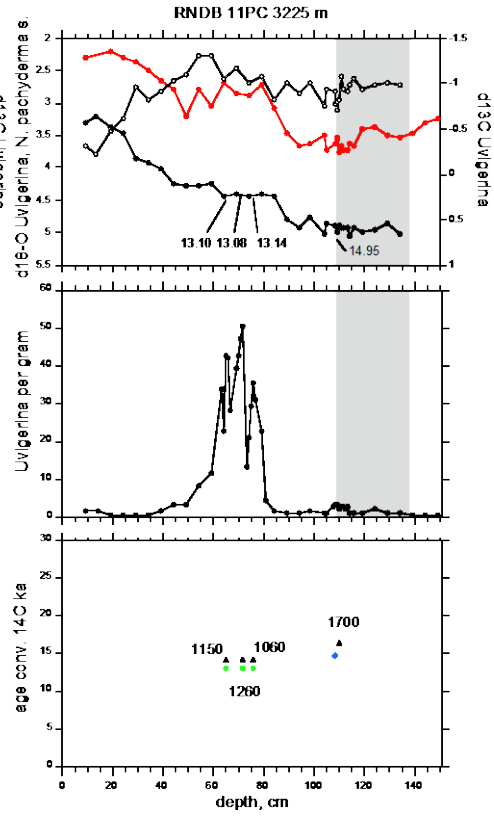
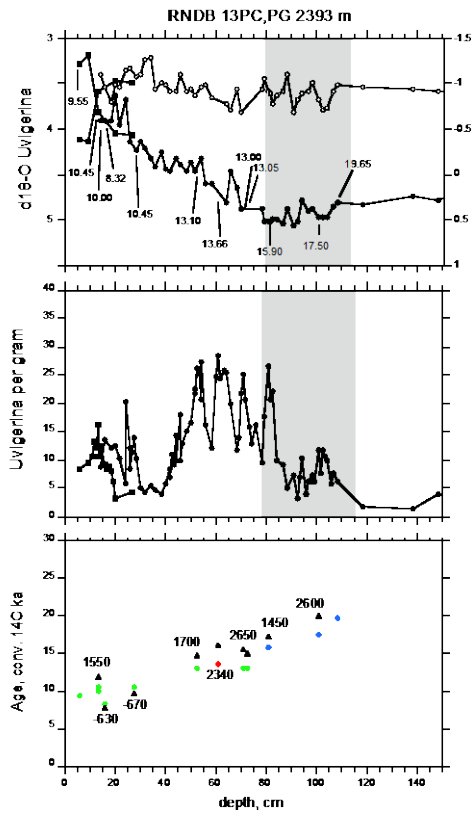
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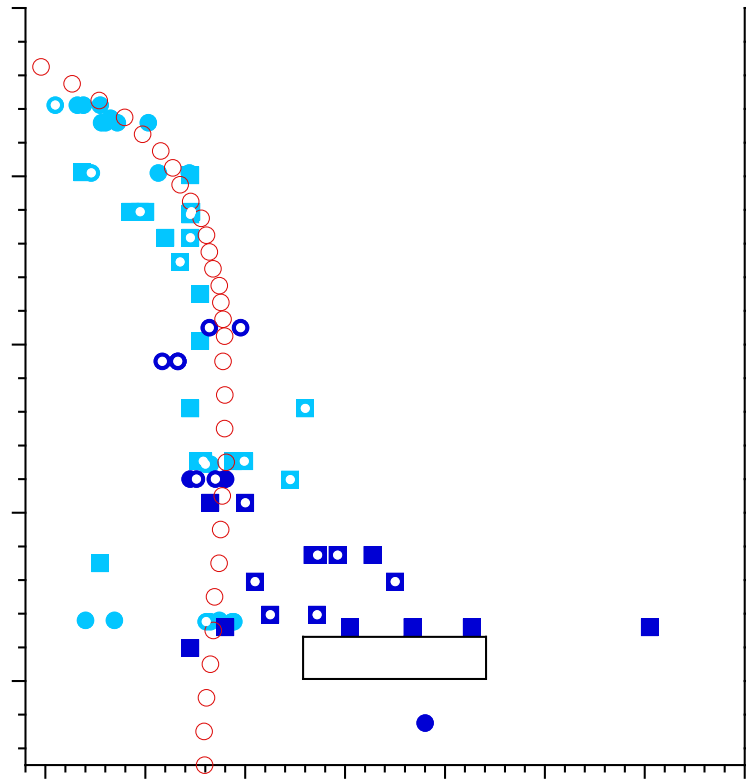
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age-400 (¹⁴C yr)

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