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Penultimate deglacial sea level timing from U/Th dating of Tahitian corals

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The timing of sea level change provides important constraints on mechanisms driving Earth's climate between glacial and interglacial states. Fossil corals constrain the timing of past sea-level due to their suitability for dating and their growth position close to sea level. The coral-derived age for the last deglaciation is consistent with climate change forced by northern hemisphere summer insolation (NHI) but the timing of the penultimate deglaciation is more controversial. Here we constrain sea level during the penultimate deglaciation, to have been ~85 m below present (mbsl) by 137 ka, and to have fluctuated on a

millennial timescale during deglaciation. This indicates that the penultimate deglaciation occurred earlier with respect to NHI than the last deglacial, initiating when NHI was at a minimum.

Corals provide a powerful archive of past sea level, but the density of coral data is biased towards sea level highstands because of the inaccessibility of fossil corals that grew during lower sea level and are now further submerged. Reconstruction of lower sea levels has relied on dredging and submersible sampling, on occasional fortuitous finds in uplifted terraces (1, 2), and on the challenging approach of coral-reef drilling. Such drilling, while technically demanding and expensive, has yielded valuable records of sea level change for the last deglacial(3, 4) and more limited constraints on the onset of the last interglacial(5). To target deeper and earlier portions of the sea level curve, IODP Expedition 310 (“Tahiti Sea level”) drilled submerged reefs in seawaters ranging from 41.7 to 117.5 m(6). The island of Tahiti Nui, French Polynesia, is located in the southern tropical Pacific, and is therefore distant from locations of glacial ice sheets. Sea level change at Tahiti during deglaciation is, therefore, dominated by the addition of melt-water to the oceans rather than the effects of ice mass redistribution and isostasy. Steady subsidence, of 0.25 mka^{-1} (4), resulting from the load of the island on the underlying oceanic plate coupled with a location distant from ice loading makes Tahiti an ideal site to reconstruct past sea levels. Pre-LGM (last glacial maximum) material was recovered at each of the three locations where Tahiti drilling was performed (Faaa, Maraa, and Tiarei)(6) (Fig. S1) and 7 separate cores have yielded pre-LGM corals suitable for U/Th dating from 113-147 mbsl.

Corals are screened for secondary calcite and aragonite by XRD (X-Ray Diffraction) and thin-section petrography. Of the 25 pre-LGM corals analysed for U-Th isotopes (7), 12 have ($^{234}\text{U}/^{238}\text{U}$)_i ($^{234}\text{U}/^{238}\text{U}$ corrected for decay since deposition) between 137-151‰, which we take as a reasonable range based on known variability of past seawater ($^{234}\text{U}/^{238}\text{U}$) during the glacial-interglacial cycle(5, 8). These 12 are considered pristine, and are discussed further here. Replicate measurements that differ significantly have been excluded from discussion (but are illustrated in Fig. 1b as small circles).

Corals of marine isotope stage (MIS) 3 age, after a correction for subsidence (0.25 mka^{-1} (4)) occur at 105-130 mbsl with ages of 29.6-33.0 ka, and are considered to be *in situ* (Table S1 & Fig. S2). They have ($^{234}\text{U}/^{238}\text{U}$)_i ratios that agree closely with one another and there is no indication of alteration. These new coral data – once corrected for subsidence – are at substantially greater depth (by ~30 m) than corals of similar age from the Huon Peninsula(9, 10) – corrected for uplift – and than $\delta^{18}\text{O}$ -based reconstructions of sea level from the Red Sea(11, 12)(Fig. 1a). This difference is too large to be accounted for by variability of the non-eustatic component of relative sea level (RSL) at far field sites (distant from ice sheets), or error in the subsidence/uplift rates.

The most plausible explanation for the discrepancy between the new Tahitian coral data and sea level reconstructions from the Red Sea and Huon Peninsula, is that they grew in deeper water. The MIS-3 corals are part of an assemblage (encrusting *Montipora* and foliaceous *Pachyseris*, thinly encrusted by coralline algae, with the absence of infilling muds) suggesting a deeper fore-reef sub-facies, and indicative of

paleo-water-depths greater than 20 m(13). Such paleodepths are consistent with sea level remaining at ≈ 80 mbsl until at least 29.6 ka, in agreement with records from the Huon Peninsula(9, 10) and Red Sea(11, 12).

Subsidence-corrected depths of two MIS 6 corals (310-M0009D-25R-2W-41,49 and 310-M0009D-25R-2W-52,55) are at 109 mbsl (Fig. 1b). The abundance of tabular *Acropora* and massive *Porites*, coupled with the presence of thick algal crusts on the upper surfaces of corals and their incorporation in a coarse sandy matrix, suggests a facies similar to “robust-branching coral” (13) and indicates a probable depth of 0-6 m for these samples. These samples therefore provide the first coral-based estimate of sea level during the fully glacial portion of MIS 6. From 153.4 ± 0.5 to 152.7 ± 0.7 ka, RSL at Tahiti was <109 mbsl and likely within 6 m of this (i.e. in the range 103-109 m). Glacial isostatic adjustment modelling has indicated that, at the LGM, relative sea level at Tahiti was 6 ± 6 m lower than the purely eustatic ice volume equivalent sea level (ESL)(14). If loading was broadly similar during MIS 6 as at the LGM, as seems a reasonable approximation, a similar discrepancy would have occurred and ESL would be 97-103 mbsl at 153 ka. It is likely that sea level fell further after 153 ka towards the penultimate glacial maximum as suggested by $\delta^{18}\text{O}$ (15) (see Fig. 1b).

Two corals are from early in the penultimate deglacial. Replicate analysis of coral 310-M0019A-29R-1W-0,4 are in agreement (136.9 ± 0.9 and 137.8 ± 0.4 ka). As with the MIS 3 and MIS 6 samples, this coral has no calcite, no aragonite overgrowths, and no signs of visible alteration. The two analyses have ($^{234}\text{U}/^{238}\text{U}$)_i ratios within error of one another and within the expected seawater range. The age of

this coral indicates that RSL must have risen to <85 mbsl by 137 ka. Another coral (310-M0019A-27R-1W-62,83) has a similar uplift-corrected depth, but younger replicate ages of 133.1, 133.2 and 134.0 ka (Fig 1b). This may reflect drowning of the reef after 137 ka with accretion unable to keep up with rising sea level across the deglaciation. Alternatively, sea level may have risen to an early highstand shortly after 137 ka and then fallen back to within 20 m of the 310-M0019A-27R-1W-62,83 sample by 133 ka. Such a millennial scale sea level fluctuation has been suggested previously based on Huon Peninsula corals (2) and Red Sea $\delta^{18}\text{O}$ (16). This interpretation is supported by the lithology and fossil assemblage of the M00019A cores. Sample 310-M0019A-29R-1W-0,4 is from a unit containing a corallgal framework dominated by *Acropora*, suggestive of shallow water-depth, while 310-M0019A-27R-1W-62,83 is from a framework of massive *Porites* indicating a water depth from 0-25 m(13) (Fig. S3). Although no U/Th ages were successful in the core section between these two dated samples, this section contains a corallgal framework of encrusting *Porites* that suggests development in deeper water during an early sea level highstand. This sedimentological evidence suggests a reversal of sea level during the penultimate deglacial, in agreement with earlier studies (2, 16). The magnitude of this early highstand – approximately two thirds of the total deglaciation - makes it substantially larger than millennial-scale variability such as that seen during MIS 3(17).

The Tahiti coral data alone indicate that global ice volume must have reduced markedly between 153 and 137 ka. A more detailed reconstruction of the deglaciation may be constructed by combining continuous sea level curves with the chronological constraints from the coral data. When comparing different localities it is important to

consider that there may be regional differences in RSL, and between RSL and ESL that strictly indicates deglaciation. RSL at far-field sites such as Tahiti provides a reasonable approximation of ESL, although the isostatic component is important at the start and end of deglaciations. RSL at Tahiti is expected to be lower by 0-15 m during the early deglacial and higher than ESL by 1.5-3.5 m (14). Tahiti RSL during the early deglacial may therefore lag ESL slightly.

To assess the magnitude and duration of the penultimate deglacial sea level change, we use the sea level reconstruction of Bintanja et al(15), based on modelling the ice volume component of deep-sea benthic $\delta^{18}\text{O}$, and drape this over our far-field coral data (Fig. 1b), maintaining the duration of the deglaciation and altering only the timing. The original timescale of the Bintanja et al record is based on the Lisiecki & Raymo(18) timescale so comparison of our timing of the deglaciation also provides some assessment of the error in that timescale. The Bintanja et al reconstruction uses a 3 ka mean of the input data so any millennial scale structure to the deglaciation is not represented. For consideration of millennial scale variability we compare the coral data to modelled sea level from $\delta^{18}\text{O}$ of planktonic foraminifera from the Red Sea(16), although that record does not extend to the glacial maximum .

Adjusting the Bintanja et al timescale to be consistent with Tahiti corals requires a shift to 4.5 ka older ages. This suggests the midpoint (the time by which sea level rose to half of the total glacial-interglacial rise) of the penultimate deglaciation occurred at ≈ 136 ka, consistent with the previous assessment from some high-stand corals(1, 20) and from U/Th dating of Bahamas sediments(19) It is ≈ 2.5 ka older than that in the Red Sea sea level curve, which has a chronology based on the highstand age in an

“open-system” coral compilation of Thompson & Goldstein(21). The timing presented here is also slightly earlier than the North Atlantic benthic $\delta^{18}\text{O}$ change placed on a chronology based on correlating ice-rafted-debris events with weak Asian monsoon intervals, precisely dated in Chinese speleothems (22) (Fig. 3).

The sea level rise at the penultimate deglacial can also be compared with the accompanying rise in atmospheric CO_2 on the independent chronology of Kawamura et al(23) for the Dome Fuji ice core record based on matching N_2/O_2 variations to local insolation at the ice core site (Fig. 2). This comparison indicates no resolvable difference in timing between sea level and CO_2 . This is in disagreement with a sea-level lag of 4 ka previously inferred by using ice-core atmospheric $\delta^{18}\text{O}$ as a lagged response to sea level change(24) This suggests that atmospheric $\delta^{18}\text{O}$ is not a reliable indicator of sea level and that the 1‰ shift seen at the deglaciation, although of similar size to that occurring in seawater during deglaciation, is controlled by changes in the Dole effect across the deglacial (25). The apparent synchronicity of sea level and CO_2 change at this deglacial (and the only slight sea level lag during the last deglaciation(3, 23)) means that mechanisms involving sea level changes as drivers of CO_2 change are no longer falsified by timing constraints, as had previously been suggested(24).

Determining the start of sea level rise is useful to elucidate the cause of the deglaciation, but is harder to define than the midpoint of deglaciation. The coral age at 137 ka constrains the start of the deglaciation to be at least that old. A more gradual start to the penultimate deglaciation, as is suggested by the Bintanja & van der Wal curve, would place the onset of ice sheet collapse at about ≈ 142 ka (Fig. 1b).

Early orbitally tuned chronologies(26, 27) followed the suggestion of Milankovitch(28), and assumed the rate of ice sheet collapse should be greatest when NHI was at a maximum because of enhanced summer melting. Depending on the precise month used to define the peak of summer, this suggests a maximum of melting rate close to 129 ka. Tahiti corals indicate that sea level is clearly rising before this at the penultimate deglacial, indicating that orbitally tuned chronologies(26) may be incorrect by up to 7 ka. It is clear that the phasing of the penultimate deglacial is very different from the last deglacial and that tuning approaches assuming a constant phasing are inappropriate if millennial scale accuracy is required.

Melting of the penultimate deglaciation starts during a minimum of NHI (Fig. 3). This requires the climate system to be unusually poised for deglaciation, if it were driven by NHI, with the subsequent rise in NHI being responsible for pacing the rate of melting throughout the deglaciation (somewhat at odds with the millennial variability seen during the deglaciation). It is interesting that the initiation of both this and the last deglaciation occur during southern hemisphere summer insolation maxima, albeit with slightly different phasing, suggesting possible role for the Southern Hemisphere. The different phasing of this deglaciation relative to the last one, however, may indicate a more complex relationship between insolation and deglaciation, perhaps involving a stochastic response to insolation (29) or control not by a single season and latitude. Any mechanism proposed to link insolation to orbital-timescale climate change must be tested against the difference in phasing of the two most recent deglaciations as constrained by U-Th dating of corals.

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Figure 1: Sea level plot for MIS 3, MIS 6 and the penultimate deglaciation. (a) Age versus subsidence-corrected depth plot showing *in situ* samples measured in this study as filled red circles. Paleo-water depth estimate of greater than 20 m is illustrated with a red bar, the dashed continuation illustrates the possibility that water depth may have been even greater. Subsidence rate used is 0.25 mka^{-1} for consistency with the previous coral record from Tahiti(4). Shown for comparison are the age vs reconstructed depth for corals from the Huon Peninsula (blue diamonds(7)). Subsidence/uplift rates are illustrated for both Tahiti and a section of the Huon Peninsula by orange lines. Sea level reconstructions from the Red Sea are shown in green(11), and purple(12). The deglacial coral sea level record from Tahiti(4) is shown in red diamonds. (b) Subsidence-corrected coral elevations versus age, for the penultimate deglaciation. New data from this study are reported as filled red circles; small circles are corals which are interpreted to have undergone some alteration, while large circles are considered to have robust chronology. Red boxes represent paleodepth estimates based on fossil assemblages and lithologies(13), the dashed boxes illustrate the large possible depth range of these facies. Existing coral data(7) are shown as open symbols. A coral from (30) is shown, green filled red square, from which the timing of MIS 5e in Lisieki and Raymo (18) is constrained. The green line is the sea level reconstruction based on an ice sheet model and high latitude air temperature(15), but with the chronology adjusted to be 4.5 ka older based on the Tahiti coral data. The blue line is a RSL reconstruction from the Red Sea (16) with the chronology adjusted +2.5 ka. The grey dashed lines are timing constraints of the deglaciation for the suggested start (142 ka) and, and the time by which sea level must have risen above 85 mbsl at Tahiti (137 ka)

Figure 2: The timing of the increase of atmospheric CO₂ and decrease of atmospheric δ¹⁸O compared to sea level rise across the penultimate deglaciation. Sea level rise is illustrated by the curves of Bintanja et al 2005(15), and Siddall et al 2006(16), adjusted to fit the new coral constraints on this study. The timing of CO₂ rise is based on the Dome Fuji record, on the time scale of Kawamura et al 2007(23). The uncertainty of this timescale, shown by the horizontal bar, is ±2.0 ka at 134 ka(23). The Vostok CO₂ time series is aligned to this Dome Fuji record to provide the timing of atmospheric δ¹⁸O (involving a shift of 1.6ka at the midpoint of the deglaciation relative to the GT4 Vostok timescale) (31). Note the clear lag between sea level and atmospheric δ¹⁸O. The grey dashed lines are the timing constraints of the deglaciation from Fig.1b.

Figure 3: The timing of the penultimate deglacial illustrated with: Bintanja et al 2005(15) (in green) - with the chronology adjusted (+4.5 ka) to match new coral data of this study (Fig. b); and the Red Sea record (16) (in dark blue) (adjusted by +2.5 ka). The Dome Fuji ice core $\delta^{18}\text{O}$ (23), and North Atlantic bottom water $\delta^{18}\text{O}$ (22) records are shown for comparison in grey and purple respectively. Local summer insolation is shown for 65°N (light blue) and 65°S (red) (32). The grey dashed lines are the timing constraints of the deglaciation from Fig. 1b. Note that deglaciation must start when NHI is at or close to a minimum.

Figure 1.

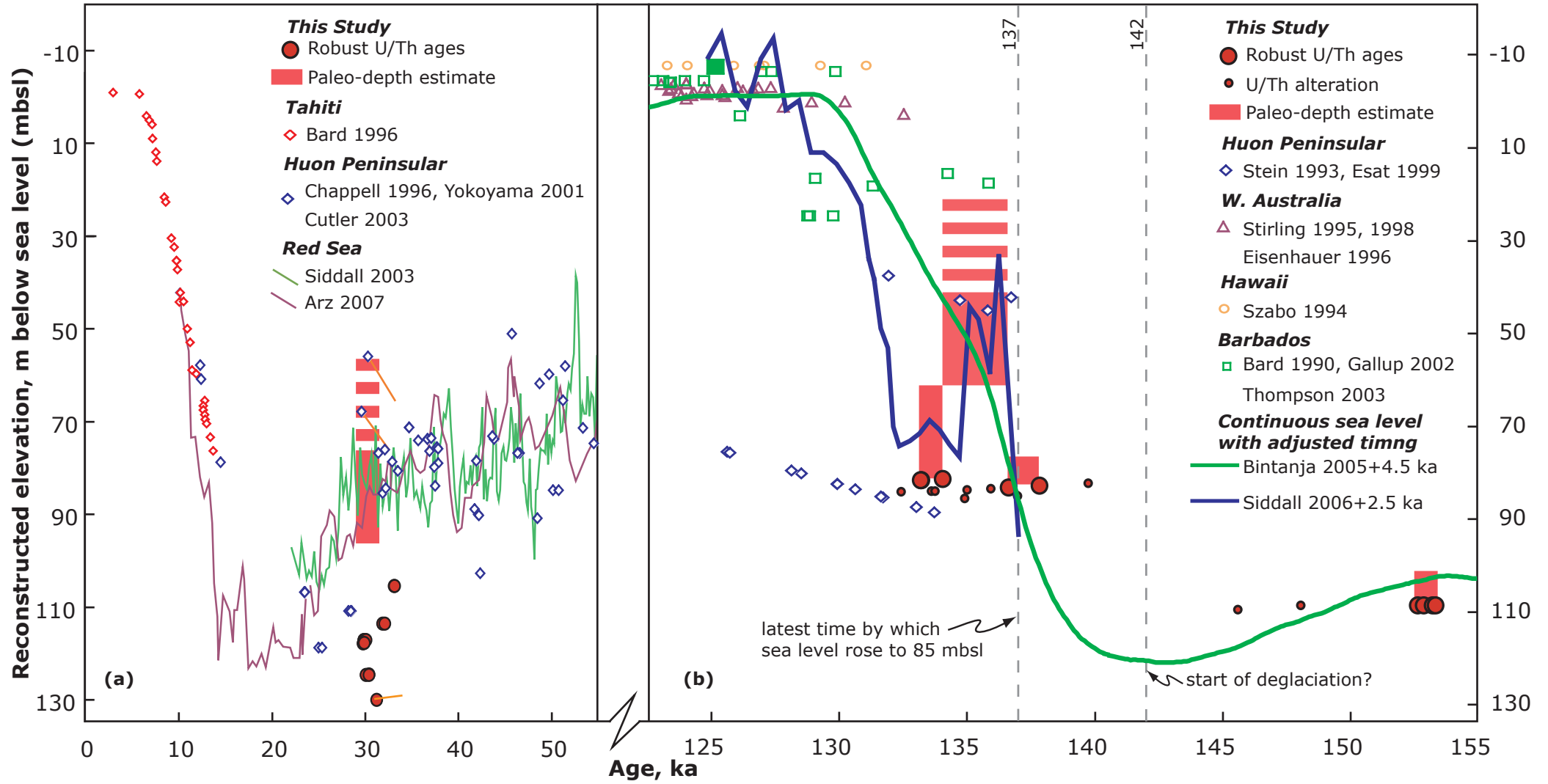


Figure 2.

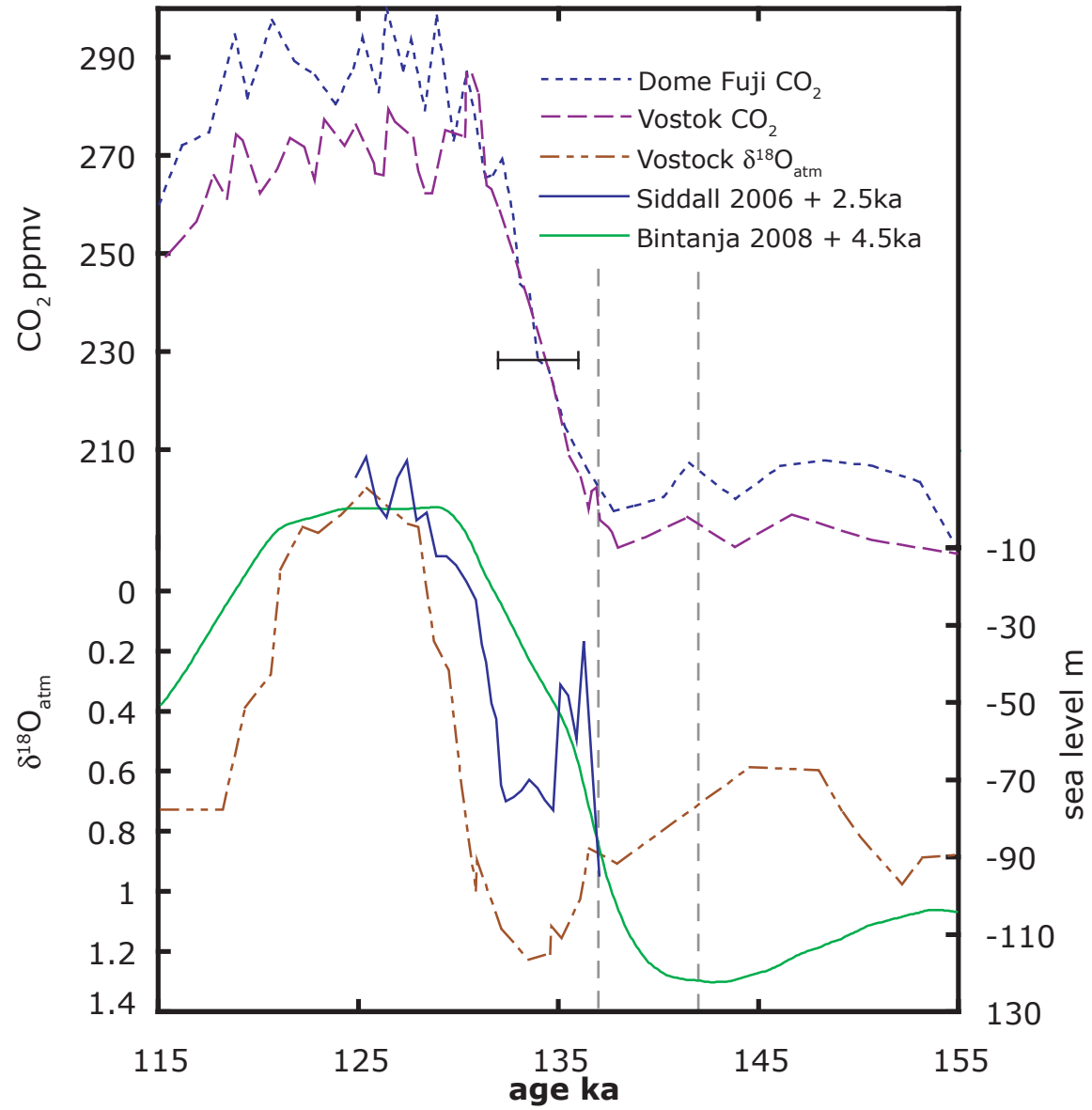


Figure 3.

