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Ice-sheet collapse and sea-level rise at the Bølling warming 14,600 years ago

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Past sea-level records provide invaluable information about the response of ice sheets to climate forcing. Some such records suggest that the last deglaciation was punctuated by a dramatic period of sea-level rise, of about 20 metres, in less than 500 years. Controversy about the amplitude and timing of this meltwater pulse (MWP-1A) has, however, led to uncertainty about the source of the melt water and its temporal and causal relationships with the abrupt climate changes of the deglaciation. Here we show that MWP-1A started no earlier than 14,650 years ago and ended before 14,310 years ago, making it coeval with the Bølling warming. Our results, based on corals drilled offshore from Tahiti during Integrated Ocean Drilling Project Expedition 310, reveal that the increase in sea level at Tahiti was between 12 and 22 metres, with a most probable value between 14 and 18 metres, establishing a significant meltwater contribution from the Southern Hemisphere. This implies that the rate of eustatic sea-level rise exceeded 40 millimetres per year during MWP-1A.

Although dynamic responses of the Greenland and Antarctic ice sheets to climate forcing may already be contributing to present-day sea-level rise[1], projections of sea-level change for the twenty-first century do not fully include potential changes in ice dynamics[2]. As acknowledged by the IPCC[3], the vulnerability of Greenland and Antarctica to ongoing warming and related discharge feedbacks remains a major source of uncertainty in projected sea-level rise[4]. Reconstructions of past sea-level changes have provided evidence for large-amplitude and rapid discharges of fresh water from continental ice sheets. Several sea-level records suggest that the glacioeustatic rise following the Last Glacial Maximum (LGM) was characterized by brief periods of extremely rapid sea-level rise[5–10]. These short-term events, referred to as meltwater pulses, probably disturbed oceanic thermohaline circulation and global climate during the last deglaciation[11,12]. The exact chronology, origin and consequences of these ice-sheet melting episodes remain unclear. But understanding these episodes is of the utmost importance when considering current uncertainty surrounding potential collapse of large ice sheets in response to recent climate change[13].

The most extreme deglacial event, MWP-1A, was initially identified in the coral-based sea-level record from Barbados[5], where a sea-level rise of ~20 m was inferred between 14,100 and 13,600 years before present (14.1–13.6 kyr BP; from here on, all ages are given as kyr before present (BP), where 'present' refers to AD 1950)[6]. However, this event remains mysterious. Several records bear witness to its occurrence[8,14,15], although no broad agreement has emerged regarding its timing. Because of this lack of consensus, the temporal relationship between MWP-1A and abrupt (millennial-timescale) climatic events that punctuated the last deglaciation are the subject of considerable debate[12,16]. Additionally, the location(s) of melting ice responsible for this prominent feature of the last deglaciation remains elusive[17].

Two conflicting scenarios have been proposed to link the timing and source(s) of MWP-1A to the climatic history of the last deglaciation. On the basis of the Barbados record's chronology[5,6], it was initially argued that this episode of rapid sea-level rise was caused by a partial melting of Northern Hemisphere ice sheets (NHIS)[5,18,19]. This 'Northern' scenario was consistent with results from a coupled ocean–atmosphere general circulation model (GCM), in which massive freshwater input to the North Atlantic would result in a weakening of the Atlantic meridional overturning circulation (AMOC) and, through the reduction of deepwater formation in the Nordic Seas, the rapid cooling of the Northern Hemisphere[11]. In this scenario, MWP-1A may have initiated the Older Dryas cold event that abruptly ended the Bølling warming about 14.1 kyr ago[14,16].

In contrast, an alternative scenario points towards an Antarctic ice sheet (AIS) as the source of MWP-1A[17,20] and suggests a causative coupling between MWP-1A and the Bølling warm period[21]. This 'Southern' scenario suggests that MWP-1A coincided with an intensification of the thermohaline circulation at the onset of the Bølling warm period[22], rather than with a slowdown during the following cold event as predicted by the 'Northern' scenario. The 'Southern' scenario was supported by output from a GCM model of intermediate complexity showing that an MWP-1A originating from the West Antarctica Ice Sheet (WAIS) may have triggered sudden reactivation of the AMOC to lead to the Bølling warming[12]. Although still contentious, this scenario solves the apparent conundrum of the Bølling warming by providing a plausible triggering mechanism for the onset of this event, traditionally considered as marking the termination of the last glacial period.

The Tahiti record

Here we report U–Th dating of coral samples collected from the Tahiti reef slope during the Integrated Ocean Drilling Program (IODP) Expedition 310, 'Tahiti Sea Level'[23]. Tahiti is a far-field site located at a considerable distance from major former ice sheets and is characterized by slow and regular subsidence rates of ~0.25 mm yr^{-1}, as consistently assessed by several approaches. Considering a total range of 0.2–0.4 mm yr^{-1} suggested by these approaches, the uncertainty on the assessment of the MWP-1A amplitude, arising from the correction of island subsidence during MWP-1A, is entirely negligible (see Supplementary Information). Previous reconstructions of the

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deglacial sea-level rise were established from holes drilled onshore through the modern barrier-reef in front of Papeete harbour[14,24]. The record was continuous from 13.9 kyr ago to present, but did not reach the critical MWP-1A period.

A specific target of Expedition 310 was the extension of the previous Tahiti sea-level record to cover earlier portions of the deglaciation. This was performed by offshore drilling of the Tahitian fore-reef slopes seaward of the present-day barrier reef (Fig. 1). These coring operations[23] recovered more than 400 m of post-glacial reef material, ranging from 122 to 40 metres below modern sea level (m.b.s.l.) in three distinct areas (Maraa, Faaa and Tiarei) around Tahiti (Fig. 1).

Our reconstruction of sea level relies on absolute U–Th dating of corals, belonging to coralgal (that is, coral and algal) assemblages indicative of a range of modern reef environments, from the shallow reef crest to the deepest reef slope. Eighty U–Th ages were determined on coral samples recovered from twenty-three holes drilled at fourteen different sites. These new data extend the Tahiti record to cover the last 16 kyr BP (Fig. 2), and provide a complete and detailed record of sea-level rise during this key period of the last deglaciation. In each hole, all of the ages are in stratigraphic order (Supplementary Fig. 2). However, even for the closely spaced holes, significant differences in recorded water depths may be observed (see, for example, the difference recorded between Site M0024 versus Site M0009 that may be up to ~10 m; Supplementary Fig. 2). The depth distribution observed for the various coral species analysed here is broadly consistent with their present-day biological zonation (Supplementary Fig. 4). The large number of holes drilled in the fore-reef slope, as well as their widespread distribution, ensured the recovery of the depth distribution of reef diversity and varying responses of reef development to sea-level rise. Our observations compare favourably with a reef accretion model[25], suggesting heterogeneous reef development induced by multiple factors including the following: spatially random (patchy) colonization; varying accretion patterns; and rugged topography of the pre-glacial surface that partially controlled the post-glacial reef initiation and growth following flooding. Our record, based on several contemporaneous cores, is therefore more representative than a record derived from a single drill hole which may provide a misleading impression of reef response to sea-level rise[25].

Sea-level rise during early deglaciation

The two oldest samples, dated at 15.74 ± 0.03 kyr BP and 16.09 ± 0.04 kyr BP, are robust branching *Pocillopora* collected at the interface of the underlying Pleistocene unit in cores 24A-15R and 9B-15R. These samples belong to a shallow-water coralgal assemblage, <10 metres water depth (m.w.d.), and indicate a Relative Sea Level (RSL) of 117–107 m.b.s.l. during that time. This RSL estimate is strengthened by the presence of an encrusting *Montipora* collected at a subsidence-corrected depth of 114 m.b.s.l. in core 25B-11R. Dated at 15.31 ± 0.02 kyr BP, this sample is associated with vermetid gastropods that are indicative of a very shallow environment (<~5 m.w.d.)[26]. From these observations, we may infer an RSL of 117–109 m.b.s.l. during the early part of the deglaciation at Tahiti (see Fig. 2).

Because of glacial isostatic adjustment (GIA), the RSL records from different sites cannot be compared directly, even in far-field regions[20]. For the time window 14–20 kyr BP, GIA models produce an RSL that is lower at Tahiti than eustatic sea level[20,27], in contrast to other sites commonly used for the analysis of sea-level change (Barbados, Bonaparte Gulf and Huon Peninsula) where GIA effects lead to local sea level lying above the eustatic value. By taking this factor into account, our 117–109 m.b.s.l. RSL estimate at 16 kyr BP is therefore in good agreement with observations from the Sunda Shelf (Supplementary Fig. 8) for the same period[8]. RSL observations from Barbados and Bonaparte Gulf display a dense cluster of samples dated at about 18–19 kyr BP, which strongly constrains eustatic sea level to a depth less than 110 m in this interval[28]. Therefore, a comparison with our data suggests that, during the early stage of deglaciation, after the MWP that occurred at 19 kyr BP[9,10], the eustatic sea level (ESL) remained stable or rose only slightly during the time span surrounding the Heinrich 1 event (probably no more than 5 m for ~3 kyr).

For the time window spanning 16.1–14.6 kyr BP, hole 24A (from the outer ridge at Tiarei) delineates the lower envelope of sea-level change. In this hole, coralgal assemblages are indicative of a very shallow environment and were able to keep pace with rising sea level during this period. The pre-MWP-1A RSL is well constrained by three coral samples collected at a subsidence-corrected depth of 105 m.b.s.l.: a massive *Montipora* sample dated at 14.65 ± 0.02 kyr BP in core 15A-37R from Maraa; and two robust branching *Pocillopora* samples dated at 14.58 ± 0.05 kyr BP and 14.61 ± 0.03 kyr BP in core 24A-10R from Tiarei (see Supplementary Information and Supplementary Fig. 3). These two last corals belong to a coralgal assemblage that typifies a shallow-water environment of less than 10 m.w.d. and are associated with vermetids that are indicative of shallow-water conditions (<~5 m.w.d.)[26]. This places a conservative constraint of 105–100 m.b.s.l. on the pre-MWP-1A sea level at 14.65 kyr BP. A moderate

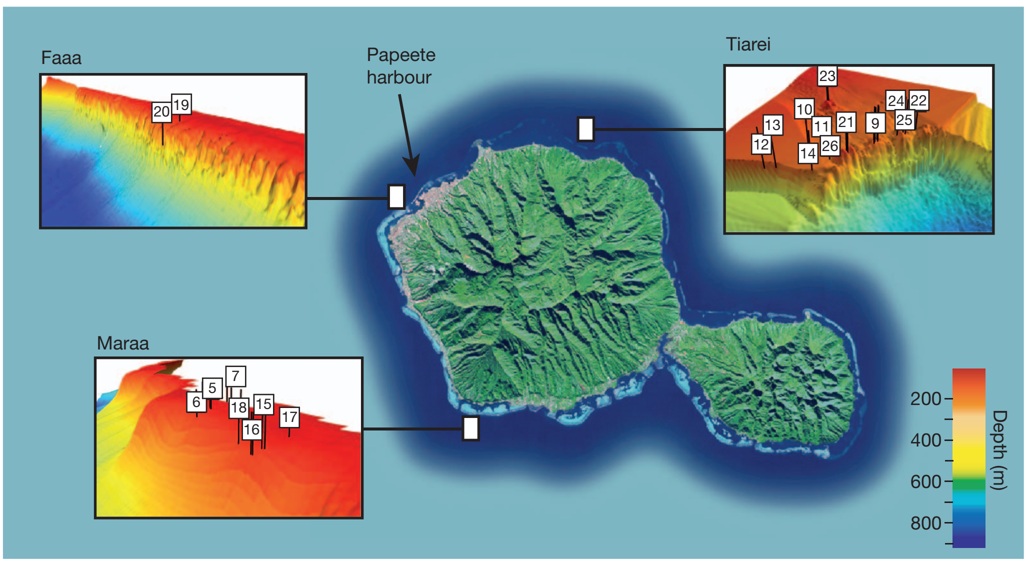

Figure 1 | A Landsat image of Tahiti island. Shown are the locations of the three areas (Tiarei, Maraa and Faaa) drilled during IODP Expedition 310, as well as Papeete harbour where onshore holes were drilled previously. A total of 37 boreholes were cored during IODP 310 at 22 different sites providing more than 400 m of post-glacial reef material[23]. Insets show the bathymetry for each site, with the location of the different drilled holes.

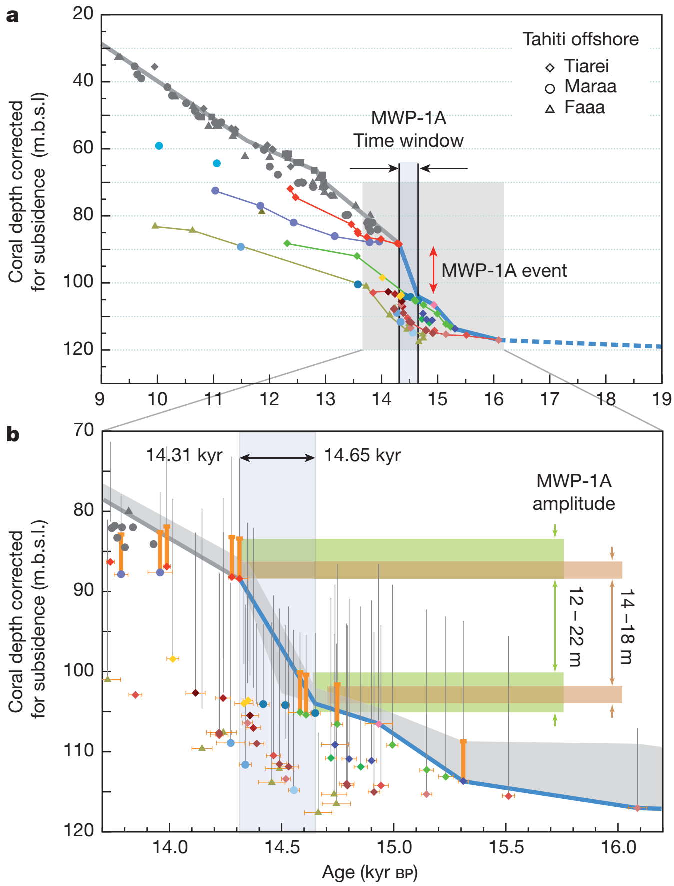

Figure 2 | The deglacial Tahiti sea-level curve. a, Sea level reconstructed from U–Th dated corals recovered in long holes drilled onshore and offshore Tahiti island. Coral depths are expressed in metres below present sea level (m.b.s.l.) and are corrected for a constant subsidence rate of 0.25 mm yr^{-1} (see Supplementary Information). Grey and coloured symbols show respectively coral samples collected in onshore holes[14,24] and in offshore holes drilled during IODP Expedition 310. Red diamonds show key samples from the inner ridge of Tiarei (Site M0023). Thick blue line shows the lower estimate of the Tahiti RSL curve (see Supplementary Information); it extends the grey curve determined by linear fits of onshore sea-level data[24] and clearly indicates the occurrence of a rapid rise of the sea level (orange arrow) related to the MWP-1A event. The shaded time window and black arrows highlight the tight chronological constraints derived for MWP-1A from the Tahiti record. **b,** Magnified view of the MWP-1A time window. The vertical grey bars reported for each coral sample correspond to their optimal bathymetric habitat range inferred from the coralgal assemblage identification (see Supplementary Information) and thick orange bars indicate samples associated with vermetid gastropods that are indicative of a shallow environment (0–5 m.w.d.). The shaded grey band illustrates our estimate of the most likely range of the Tahiti RSL over the last deglaciation. The ranges of uncertainty estimated from the bathymetric range of coralgal assemblages for the pre- and post-MWP-1A sea-level positions are illustrated by the horizontal green bands. The resulting extreme bounds for the MWP-1A amplitude (12 and 22 m) are also indicated (green bands and arrows). Several arguments given in the Supplementary Information suggest that these conservative estimates can be trimmed to 14 and 18 m (brown bands and arrows). Thick blue line and thick grey line are as in **a**.

sea-level rise of 4–14 m is therefore inferred for the period from 16.1 to 14.65 kyr BP.

The earliest bound for the initiation of the MWP-1A jump of sea level is probably within the time range given by those three samples (14.58–14.65 kyr BP). Moreover, the two *Pocillopora* samples dated at 14.58 and 14.61 kyr BP could have already grown at a reasonable water depth (up to 5 m.w.d.). Thus, they may have already accommodated a part of the sea-level rise related to MWP-1A, implying that the inception of MWP-1A could have occurred somewhat earlier (see the upper bound of the shaded grey area in Fig. 2; see also Supplementary Fig. 4). The maximum age for the onset of MWP-1A could thus be close to the oldest of these three corals, dated at 14.65 kyr BP. It must be emphasized that this only provides us with the uppermost limit for the onset of MWP-1A, and we cannot rule out that the jump may have started significantly later, as young as 14.5 kyr BP, as potentially marked by massive *Montipora* samples of core 15A-36R that characterize a shallow environment (see Supplementary Figs 3–5).

Occurrence of MWP-1A

The occurrence of MWP-1A is revealed by a major discontinuity in the upper envelope of the data points in the new Tahiti RSL record (Fig. 2). The next shallowest *in situ* samples in the sequence are two branching *Pocillopora* dated at 14.28 ± 0.02 kyr BP and 14.31 ± 0.04 kyr BP in cores 23B-12R and 23A-13R (see Supplementary Information and Supplementary Fig. 3). These coral samples, recovered at a subsidence-corrected depth of 88 m.b.s.l., are the first datable corals, showing clear evidence of an in-growth position, to colonize the pre-glacial substratum after the MWP-1A sea-level jump. These samples are critical, as they provide the most robust constraint on MWP-1A timing and clearly indicate that the sea-level jump was complete before 14.31 kyr BP. These data lie on the extension of the general trend depicted by onshore holes[14,24] (Fig. 2) and highlight a regular, slow rate of sea-level rise after MWP-1A. These corals are associated with vermetids, thus indicating a very shallow environment (<~5 m.w.d.). We infer a conservative estimate of 88–83 m.b.s.l. for the post-MWP-1A sea level.

The MWP-1A event also coincides with a major change in reef development strategy, as illustrated by numerous samples dated in all drill holes collected on the outer edge of the fore-reef slopes. Before MWP-1A the reef kept pace with sea level, whereas a widespread deepening and backstepping occurred after MWP-1A. This change in reef response is coincident with changes in the coralgal assemblage composition, such as in Hole M0024A (see Supplementary Information and Supplementary Fig. 3), where shallow-water assemblages—dominated by robust branching *Pocillopora*, massive *Porites* and encrusting *Montipora*—change to branching *Porites* species, which typify an environment characterized by moderate energy and light intensity.

General features of reef geometry can be simulated with a two-dimensional growth model[14]. This model simulates the overall deepening of the reef sequence that follows occurrence of a rapid sea-level rise and clearly indicates that only holes drilled in the intermediate position between the outer ridge and the modern barrier reef are capable of capturing the sea-level position immediately following MWP-1A (see Supplementary Figs 9 and 10). This result probably explains the difficulty encountered by previous onshore or offshore drilling programmes (Tahiti or Barbados) to collect shallow-species coral samples that document precisely the end of MWP-1A. The IODP Mission Specific Platform overcame this difficulty by specifically targeting the reef structures located in intermediate position between the fore-reef slope and the present barrier reef, especially at the Tiarei site.

Amplitude and duration of MWP-1A at Tahiti

On the basis of the most conservative estimates deduced above for the pre- and post-MWP-1A sea level, we infer an amplitude of 17 m for the sea-level jump, with lowest and uppermost bounds of 12 and 22 m. Several arguments, discussed in detail in Supplementary Information (Supplementary Fig. 6), suggest that this range may reasonably be narrowed down to 14–18 m, with a median value of 16 m.

In view of the lower and upper limits of the MWP-1A chronozone (14.31 kyr BP and 14.65 kyr BP, respectively), the longest possible duration of the jump is ~350 years (Fig. 3). Considering the median value of 16 m for the local amplitude of MWP-1A at Tahiti, we infer an average RSL rate of ~46 ± 6 mm yr^{-1} at Tahiti. However, owing to the age uncertainty associated with its inception and termination (see

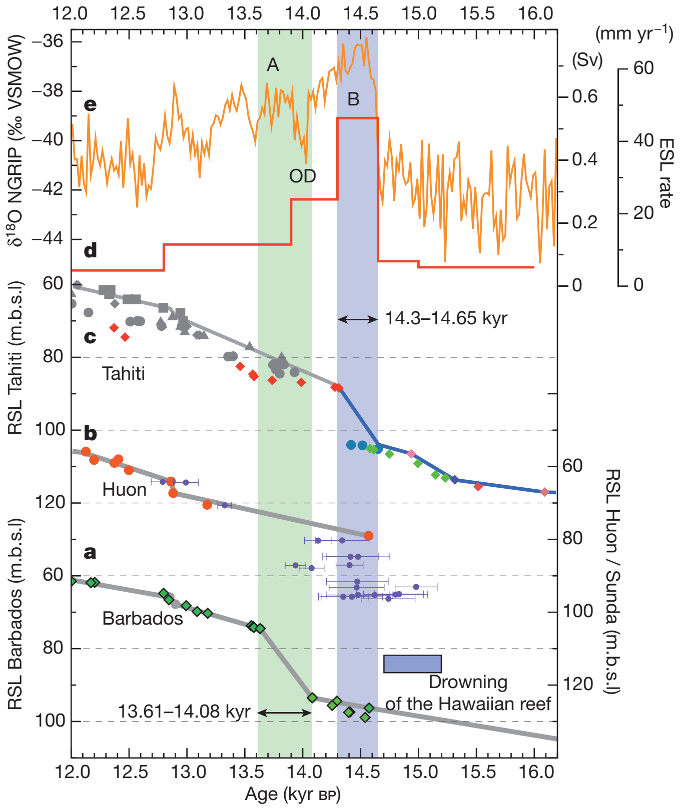

Figure 3 | Relative sea-level (RSL) records over the time window 16.5 to 12.0 kyr BP. **a**, Barbados RSL record based on U–Th dated corals (mainly *Acropora palmata*)[19,29]. The shaded green vertical band highlights the MWP-1A time window inferred from the Barbados record[19,29] **b**, Pacific RSL records (right-hand vertical axis). Red circles, Huon Peninsula record[31,32] (Papua New Guinea) based on U–Th dated corals. Purple points, Sunda Shelf record[8] based on ^{14}C-dated organic material found in sediment cores (recalibrated using IntCal09[30]; plotted errors are 1σ). The blue rectangle indicates the drowning of a Hawaiian reef 14.7 kyr ago[15]. **c**, Tahiti RSL record based on U–Th dated corals collected in holes drilled onshore (grey symbols)[14,24] and offshore (coloured symbols, this study). The shaded purple vertical band highlights the MWP-1A time window inferred from this study. **d**, Rate of glacial meltwater discharge (expressed in mm yr^{-1} and Sv, right-hand vertical axes) derived from the eustatic sea level curve determined by the GIA model (see Supplementary Information and Supplementary Fig. 11) adjusted to account for the newly obtained timing and magnitude of MWP-1A from Tahitian sea-level observations. **e**, δ^{18}O record of the North Greenland Ice Core Project (NGRIP) core plotted on its most recent timescale[34]; B, Bølling; OD, Older Dryas; A, Allerød. All depths have been corrected for subsidence (Tahiti) and uplift (all other sites) as described in ref. 24. For Tahiti and Barbados records, only samples that delineate the upper envelope are shown. Grey lines correspond to linear fits of sea-level data[24]. Greenish and bluish shaded time windows correspond to MWP-1A chronozones inferred from the Barbados record and the Tahiti record (Fig. 2), respectively.

Supplementary Information), the MWP-1A duration could have been even shorter than this estimate. An extremely sharp meltwater outburst, of the order of a century or less, is thus possible, in which case the 46 mm yr^{-1} rate of sea-level rise must be considered as a minimum value.

Timing of MWP-1A

The new MWP-1A chronozone inferred from the extended Tahiti record (that is, 14.65–14.31 kyr BP or shorter, Fig. 3) does not overlap with that previously proposed on the basis of the Barbados record (14.08 ± 0.06 to 13.63 ± 0.03 kyr BP, using the most recent updated data set[19,29]; see Supplementary Information for a full discussion of this issue).

Several other lines of evidence also suggest that MWP-1A was significantly older than suggested by the Barbados record and, ultimately, concurrent with the Bølling warming. Additional evidence comes from the Sunda Shelf sea-level record, derived from mangrove organic material collected from a shallow siliciclastic platform[8]. This record shows a very sharp sea-level rise dated at a conventional ^{14}C age of 12.42 ± 0.06 kyr BP (1 s.d., $n = 17$; Supplementary Fig. 8) coinciding with the 500-year-long ^{14}C plateau that encompasses the Bølling period. Using the IntCal09 calibration curve[30], the mean calendar age of the MWP-1A event recorded on the Sunda Shelf can be refined to 14.94–14.14 kyr cal. BP (2σ interval, see Supplementary Information for more details regarding this age calculation).

The revised MWP-1A timescale inferred from the new Tahiti record is also coherent with the recent extension of the Huon Peninsula record[31,32], where the oldest sample of the post-glacial reef sequence dated at 14.56 ± 0.05 kyr BP places an upper constraint on the end of MWP-1A (Fig. 3). Further indirect evidence is provided by the drowning of coral reefs offshore from Hawaii, which occurred at 14.7 kyr BP and has been proposed to be caused by a dramatic increase in sea level related to MWP-1A[15].

These records are consistent enough to revise the onset of MWP-1A so it is 500 years earlier than the date inferred from the Barbados data. Within this revised timeframe, MWP-1A can no longer be advocated as the trigger for the Older Dryas cooling event that terminated the Bølling period, as proposed previously[14,16,33]. Instead, MWP-1A coincided with the inception of the Bølling period (Fig. 3), which has been independently constrained by the GICC 05 Greenland ice core chronology at 14.640 kyr BP (with a maximum counting error of 0.186 kyr)[34]. The Tahiti record is thus compatible with the idea of a temporal relationship between MWP-1A and Bølling warming. This hypothesis is further substantiated by the concurrent occurrence of rapid flooding on shelf margins and an increase in sea surface temperature in the South China Sea at the Bølling transition[21].

Source of MWP-1A

Because they account for more than 80% of total sea-level rise during the last deglaciation, NHIS, and especially the Laurentide Ice Sheet (LIS), have commonly been considered as the sole sources for MWP-1A[5,35]. But arguments for such an LIS source faced serious objections, and led to the proposal[17] of an alternative scenario in which a significant fraction of the melt water came from Antarctica.

Direct evidence in favour of a Northern or Southern Hemisphere source remains equivocal. Most robust arguments supporting an Antarctic contribution were provided by GIA models[20,36,37]. Fingerprinting model experiments demonstrated that comparison of the size of the MWP-1A sea-level rise observed at several sites could provide helpful information about the source(s) of melting ice[36]. Predictions provided in ref. 36 showed that, when melting ice originated exclusively from the LIS, the amplitude of MWP-1A predicted for Barbados should be significantly lower than for far-field sites. This scenario predicted the greatest difference in amplitude between Barbados and Tahiti, with a sea-level rise at Tahiti almost twice that at Barbados[36].

The amplitude of MWP-1A that we assess at Tahiti (16 m) is comparable to that observed at Sunda (~16 m)[8]. At Barbados, the amplitude of the jump must be reassessed on the basis of the re-evaluation of the MWP-1A chronozone (Supplementary Fig. 7). By extrapolating the linear trend defined by hole 12 (Supplementary Fig. 7), we roughly estimate a ~15 m amplitude of sea-level rise at Barbados. The amplitudes of MWP-1A recorded at these three far-to-intermediate-field sites are thus approximately the same. Following the predictions of ref. 36, our results seem to preclude a sole LIS contribution to MWP-1A and confirm the preliminary conclusions[36] based solely on the Sunda and Barbados records. On this basis, the Barents and Fennoscandian Ice Sheets can also be considered as possible candidates for the freshwater source (see figure 2 in ref. 36), but there are several counterarguments to these ice sheets as the major sources of fresh water[17]. All other scenarios that provide equal amplitudes of MWP-1A sea-level rise require a significant Antarctic contribution.

These arguments in favour of a contribution from the AIS were reinforced by GIA predictions[20]. Those predictions showed that the optimal deglacial scenario to fit RSL observations at Barbados, Tahiti, Huon Peninsula and Sunda Shelf during late glacial time required a MWP-1A with a total amplitude of 23 m, which included an AIS contribution of 15 m with a total NHIS contribution of 8 m (6 m from the LIS).

Using a realistic GIA model (see Supplementary Fig. 11 and Supplementary Information), which uses the Earth model proposed in ref. 20, we performed a new set of simulations that agree well with the conclusion of Bassett *et al.*[20], pointing towards a substantial contribution from the AIS. It is difficult at this stage, however, to conclusively determine the relative contributions of NHIS and the AIS to MWP-1A because these approaches (fingerprinting and more general GIA modelling) are hampered by uncertainties surrounding the MWP-1A-induced relative sea-level amplitude, especially at the intermediate-field site of Barbados. Following previous studies[27,36], which conclude that the MWP-1A amplitude recorded at Tahiti is amplified by 10–30% with respect to its eustatic amplitude, our results are consistent with a eustatic MWP-1A rise of roughly ~14 m during the time window 14.65–14.3 kyr BP, leading to a rate of eustatic sea level rise of 40 mm yr^{-1}. Note that this value is significantly lower than the 20–25 m of eustatic rise often reported in the literature[20,36]. Considering the growing body of evidence[20,36,37] that suggests that a substantial fraction of MWP-1A originated from Antarctica, it is probable that the AIS contributed at least half of the ~14 m eustatic sea-level rise observed during this event. It is worth noting that this estimate of the Antarctic contribution allows us to balance the freshwater budget required for MWP-1A, taking into account NHIS contributions that have been independently assessed to be between 5 and 10 m of sea-level equivalent ice volume[38,39]. Recent estimates of AIS contribution to the last deglaciation indicate that its contribution was <20 m and perhaps lower than 10–15 m (refs 40–42), implying that a significant, if not the major, part of the AIS contribution to the last deglaciation occurred during MWP-1A.

Implications of the revisited MWP-1A history

The IODP Expedition 310 provides significantly improved constraints on the timing of MWP-1A, demonstrating that MWP-1A ended before 14.3 kyr BP and that it started after 14.65 kyr BP. This makes MWP-1A coeval with the Bølling warming, suggesting a temporal, and probably causal, relationship between these two prominent deglacial features. Owing to the dating uncertainty of the Bølling inception in the Greenland ice record (14.642 kyr BP with a maximum counting error of 186 years; ref. 34), it remains difficult to unravel the phasing and causal mechanisms linking—through specific atmospheric and oceanic responses—the resumption of the AMOC during the Bølling warming[22] and massive meltwater discharges in both hemispheres. Two end-member scenarios that warrant further investigation can be put forward, however:

The first scenario is that proposed in ref. 12, based on GCM simulations showing that a rapid freshwater discharge originating from the AIS could have led to an intensification of the AMOC. The associated northward ocean heat flux would trigger the Bølling warming in the Northern Hemisphere and a rapid melting of the LIS. But subsequent studies (for example, ref. 43 and references therein) that have tested the scenario of ref. 12 showed that the meltwater discharge may have led to competing mechanisms, enhancing or weakening the AMOC, which collectively lead to a subdued climatic response in the Northern Hemisphere[43].

In the second scenario, the phasing of events is reversed, with an initial AMOC increase and associated northward ocean heat transport causing the Bølling warming, which led to rapid melting of NHIS, in particular the LIS. The resulting sea-level rise drove in turn a dramatic collapse of the AIS. Indeed, the WAIS was partly marine-based during the LGM and thus probably sensitive to the break-up and loss of buttressing ice shelves. In any case, most of the WAIS is characterized by unstable conditions, with bedrock below sea level and slopes downward from the margins towards the interior[13].

In fact, these two scenarios are not mutually exclusive and could have acted in concert during the MWP-1A chronozone, reinforcing each other. They are both compatible with our sea-level and source fingerprinting study, which implies that meltwater injections forming the MWP-1A event originated from ice sheets in both Antarctica and the Northern Hemisphere, including the LIS. In principle, meltwater injection into the North Atlantic could have counteracted the AMOC increase, but the strength of this negative feedback depends on the exact location and mode of meltwater release. Several studies suggested that LIS meltwater was funnelled through the Mississippi drainage system, before being released in the Gulf of Mexico as a hyperpycnal flow[38,44], with a negligible impact on the AMOC[39,45,46].

The two scenarios have similar ingredients but differ in their ultimate trigger, AIS collapse or AMOC increase. These abrupt events could be linked to threshold responses to the gradual warming of the Southern Hemisphere that occurred under external forcings (orbital and greenhouse-gas changes) during the early part of the deglaciation[47,48].

Much research remains to be done to document the precise sequence of events during the MWP-1A chronozone. This will come from coring coral reefs at other sites (for example, Barbados and the Seychelles[27]), from study of open-ocean sediments in the vicinity of former ice sheets, and from modelling work to simulate the complex interplay between ice sheets, ocean and atmosphere. Whatever the causes that led to the MWP-1A event and the Bølling warming, and despite the fact that the total eustatic magnitude of this event is reduced compared to previous estimates, our results prove the existence of a dramatic collapse of past ice sheets at a eustatic rate exceeding 40 mm yr^{-1}, with a substantial contribution from Antarctica. We note that this rate is at least four time as large as the average rate of deglacial sea-level rise of ~10 mm yr^{-1}; see ref. 24 and Supplementary Information. Understanding this singular event will shed light on the dynamical behaviour of large ice sheets in response to external forcing or internal perturbation of the climate system. This topic is crucial in the context of the present warming, as modern ice sheets have been shown to be contributing directly to the recent acceleration in sea-level rise[1,2].

METHODS SUMMARY

Before U–Th dating, rigorous mineralogical and isotopic screening criteria were applied to discard coral samples that suffered any post-mortem diagenetic alteration of their aragonite skeleton. In particular, using X-ray diffraction[30], we made an effort to improve the detection and quantification of a very small amount of secondary calcite. Coral samples showing a calcite content of more than 1% were discarded. Most of the U–Th analyses were performed using a VG-54 thermo-ionization mass spectrometer equipped with a 30-cm electrostatic analyser and a pulse-counting Daly detector at CEREGE (see Supplementary Information for data and analytical issues). The initial $(^{234}U/^{238}U)_0$ values calculated for post-glacial samples yielded a mean value of 1.1458 ± 0.0020 (2σ), falling within the most recent determinations of modern sea water and corals[49]. Additionally, for corals of the same age, $(^{234}U/^{238}U)_0$ values were highly consistent (that is, within an analytical uncertainty determined for the entire course of the study of 0.8‰, 2σ), and within the larger range adopted[30] as an isotopic screening criterion in the interval 0–17 kyr BP ($(^{234}U/^{238}U)_0 = 1.1452 \pm 0.0048$, 2σ). The clustering of $(^{234}U/^{238}U)_0$ values determined in this study substantially narrows the uncertainty for the evolution of the seawater value through time compared to previous data sets (Vanuatu, Papua New Guinea and Barbados) that have encompassed the last deglaciation, highlighting the outstanding quality of the coral samples recovered in Tahiti offshore holes. Complementary and duplicated analyses were also performed by Multi-Collector Inductively Coupled Mass Spectrometry[50] and show a general good agreement within measurement uncertainties.

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