Synchroneity of meltwater pulse 1a and the Bølling warming: New evidence from the South China Sea

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ABSTRACT

A twofold decrease in long-chain *n*-alcane (*n*-nonacosane) concentrations in a downcore record from the northern South China Sea indicates a rapid drop in the supply of terrigenous organic matter to the open South China Sea during the last deglaciation, paralleled by an equally rapid increase in sea-surface temperatures, corresponding with the Bølling warming at 14.7 ka. The sudden drop in terrigenous organic matter delivery to this marginal basin is interpreted to reflect a short-term response of local rivers to rapid sea-level rise, strongly implying that the Bølling warming and the onset of meltwater pulse (MWP) 1a are synchronous. This phase relation contrasts with the widely cited onset of this MWP 1a ca. 14 ka, and implies that previous studies postulating a weakening of deep-water formation in the North Atlantic due to massive meltwater discharge during MWP 1a need to be reevaluated.

Keywords: sea level, organic geochemistry, meltwater, South China Sea.

INTRODUCTION

Following the last glaciation, sea level rose extremely rapidly (13.5–24 m in <290–500 yr) during meltwater pulse (MWP) 1a (Fairbanks, 1989; Bard et al., 1990; Blanchon and Shaw, 1995; Hanebuth et al., 2000). Together with the Bølling warming, an abrupt rise in Northern Hemisphere air temperature of at least 5 °C within a few decades during the last deglaciation, MWP 1a arguably represents the most dramatic event in Earth's climate history during the past 25 k.y., and determining its precise age is of utmost importance for a mechanistic understanding of the oceanographic, glaciological, and climatic changes that occurred during the last deglaciation (Clark et al., 1999). Nevertheless, the absolute timing of MWP 1a and its phase relation with the Bølling warming are still debated. According to the widely accepted chronology and correlation provided by Bard et al. (1996), MWP 1a ca. 14 ka corresponds to the Older Dryas, i.e., to the first major cooling event following the Bølling warming. This would imply a significant weakening of the thermohaline circulation and its associated heat transport to the North Atlantic region due to the freshwater input. Various modeling studies (Stocker et al., 1992; Manabe and Stouffer, 1997) appear to corroborate a causative coupling of MWP 1a and the Older Dryas, indicating a cessation of deep-water formation in the North Atlantic in response to a massive meltwater input. However, Lohmann and Schulz (2000) suggested that many previous models underestimated overflow over the Greenland-Scotland Ridge, and showed that meltwater discharge and continued deep-water formation in the North Atlantic can be reconciled using a new model that allows for significant deep-water formation in the Greenland-Iceland-Norwegian Sea. Furthermore, a modeling study by Clark et al. (2002a) suggests that most of the meltwater during MWP 1a originated from Antarctica, and not, as previously thought, from the Laurentide ice sheet. However, the dating of MWP 1a provided by Hanebuth et al. (2000) shows that the major rise of sea level occurred between 14.7 and 14.3 ka, i.e., synchronously with the Bølling warming in Greenland ca. 14.7 ± 0.3 ka (Stuiver and Grootes, 2000). This, in turn, would suggest that MWP 1a coincides with intensifying thermohaline circulation during the Bølling warming (Clark et al., 2002b) rather than a slowdown, as postulated earlier (Bard et al., 1996). Here we provide an independent means of establishing the phase relation between MWP 1a and the Bølling warming, circumventing the inherent uncertainties of quasi-absolute chronologies and of comparing independently dated records, e.g., ice core with coral reef (Bard et al., 1996) or siliciclastic shelf (Hanebuth et al., 2000) records.

The rapid rise in sea level during MWP 1a led to a rapid retreat of shorelines around the world (including river mouths) and a sudden flooding of large parts of exposed shelf areas, and therefore the inundation of the lower reaches of rivers close to or at the shelf margins. This rapid transgression of large, shallow areas, most notably in western Pacific marginal seas, instantaneously provided a rapidly landward-extending and deepening accumulation space for terrigenous sediment. These conditions preferentially favored deposition on the shelves and, in turn, led to an abrupt decrease in the supply of sediment to the outer shelves, the continental slopes, and the deep sea.

RESULTS

Terrestrial organic matter supply to the northern South China Sea (site 17940, 20°07′N, 117°23′E) decreased rapidly at the end of the last glaciation (Fig. 1), a change that is paralleled by a similar abrupt increase in sea-surface temperature (SST) of ~1 °C. An identical synchroneity between the drop in terrigenous sediment supply to the open South China Sea and a significant warming at the end of the last glaciation is also observed at a site from the southern South China Sea in front of the Sunda Shelf (site 17964, 06°09′N, 112°13′E; Pelejero et al., 1999a), as well as in two lower resolution cores from the open South China Sea (core 17954, 14°48′N, 111°31.5′E, and core 17961, 08°30′N, 112°20′E; Kienast et al., 2001a). This drop in terrigenous supply to the South China Sea is not the result of decreased summer monsoonal precipitation in the region, which could equally lead to reduced fluvial sediment discharge. On the contrary, a number of studies show that summer monsoonal precipitation increased during the deglaciation (L. Wang et al., 1999; Y. Wang et al., 2001), which would have led to an increase in terrigenous sediment supply to the South China Sea. Furthermore, this decrease in terrigenous organic matter supply to the South China Sea could not have been caused by a decrease in eolian transport, because the decrease in winter monsoonal strength during the last deglaciation occurred significantly prior to the abrupt deglacial warming (Wang et al., 1999).

The rapid increase in SST at two sites in the southern South China

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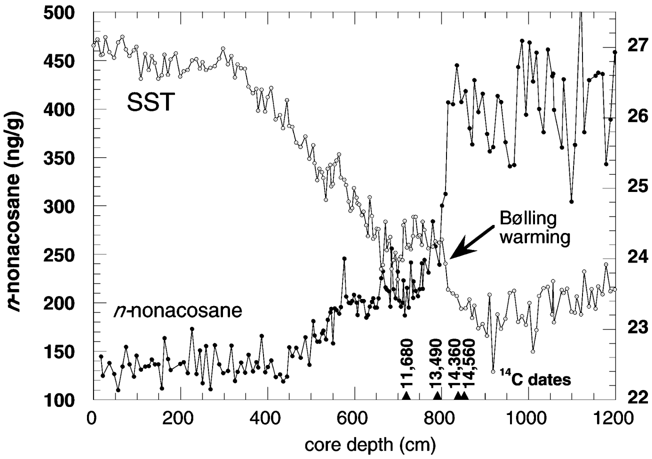

Figure 1. Alkenone ($U^{K'}_{37}$) sea-surface temperature (SST) estimates (open circles; from Pelejero et al., 1999b) and *n*-nonacosane (C_{29} *n*-alkane) concentrations (filled circles; ng/g; see Kienast et al., 2001a, for analytical method) of core 17940–2 vs. depth. Long-chain *n*-alkanes with predominance of odd-carbon-numbered members are uniquely produced by terrestrial plants, and are thus widely used in paleoceanographic studies to monitor terrigenous organic matter input to marine sediments. For simplicity, only *n*-nonacosane data are reported here. This approach is justified by uniform distribution of homologs ranging between C_{23} and C_{33} in South China Sea sediments, and close linear correlation between *n*-nonacosane abundances and total odd-carbon-numbered long-chain *n*-alkanes in selected samples analyzed for complete set of long-chain *n*-alkanes. Selected accelerator mass spectrometer radiocarbon dates bracketing Bølling warming and meltwater pulse 1a (at 722.5, 792.5, 842.5, and 852.5 cm) are from Wang et al. (1999) (see text). Note sample to sample synchroneity of sharp SST increase and *n*-nonacosane decrease at onset of Bølling warming.

Sea (core 18252-3, 9°14′N, 109°23′E and core 18287-3, 5°39′N, 110°39′E) is delimited by accelerator mass spectrometer (AMS) radiocarbon dates to have started at 14.7 ± 0.3 ka (Kienast et al., 2001b), synchronously with the Bølling warming in the Greenland GISP2 ice-core record. Similarly, we argue that the rapid warming at the end of the last glacial observed at site 17940 in the northern South China Sea corresponds to the Bølling warming. The AMS radiocarbon dates (Wang et al., 1999; Fig. 1) yield an interpolated age of the midpoint of this warming step in core 17940-2 of 15,970 +285/−260 yr. Given the unlikelihood of such a significant lead of the Bølling warming in the northern South China Sea compared to two SST records from the southern South China Sea (Kienast et al., 2001b), we attribute this apparent offset in absolute ages to locally increased reservoir ages of as much as 1200–1300 yr at site 17940, possibly caused by advection of Pacific Intermediate Waters (Wang et al., 1999). Three lines of evidence are adduced to corroborate this assertion. First, the increases in local reservoir age at site 17940 are within the range of similar variations reported from the North Atlantic (Waelbroeck et al., 2001) and the Mediterranean (Siani et al., 2001); this suggests that variations of local reservoir ages of as much as 1300 yr are not necessarily restricted to strong upwelling centers. Second, the sample to sample correspondence between the SST increase during the Bølling warming and a planktonic foraminiferal $\delta^{18}O$ decrease observed in the southern South China Sea (cores 18252-3 and 18287-3; Kienast et al., 2001b) is also evident in core 17940-2 (Pelejero et al., 1999b). Third, the detailed phasing of SST and benthic $\delta^{18}O$ from within the same core is identical in cores 17940-2 and 17964-3 from the northern and southern South China Sea, respectively (Fig. 2). Taken together, this evidence argues against a diachrony of the rapid deglacial warming step within the South China Sea.

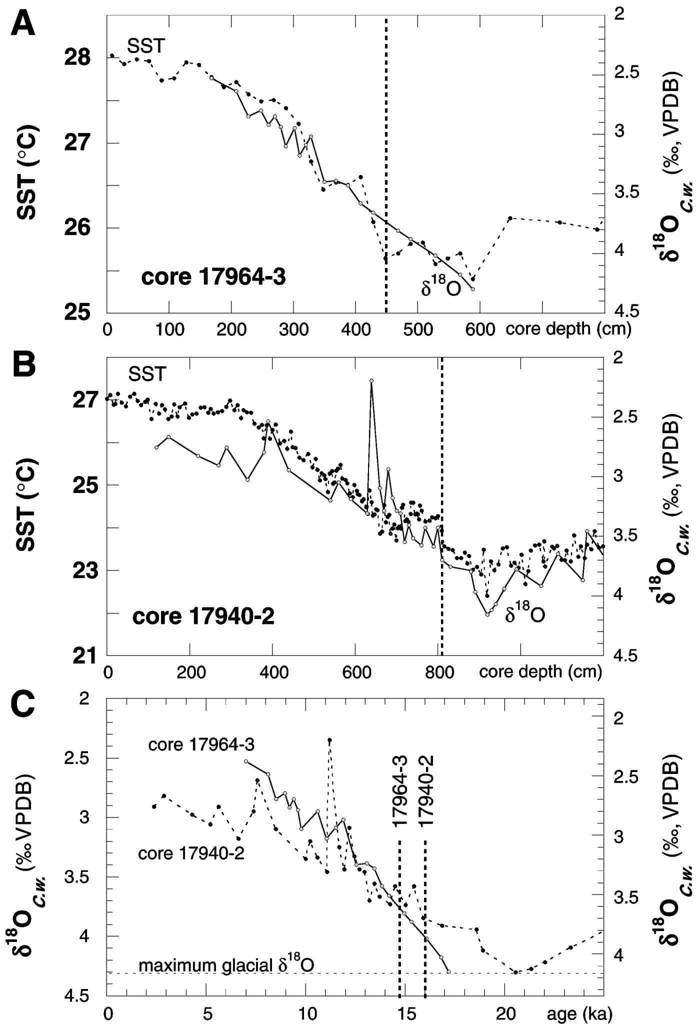

Figure 2. Alkenone sea-surface temperature (SST; solid dots, stippled line, from Pelejero et al., 1999b) and benthic (*Cibicidoides wuellerstorfi*) oxygen isotope ($\delta^{18}O_{C.w.}$; open dots, solid line, from Wang et al., 1999; VPDB is Vienna peedee belemnite) records of cores 17964-3 (A) and 17940-2 (B) versus depth from southern (06°09′N, 112°13′E, 1556 m water depth) and northern (20°07′N, 117°23′E, 1727 m water depth) South China Sea (SCS), respectively. Stippled vertical lines mark onset of abrupt deglacial warming step. Note that in both cores $\delta^{18}O$ maximum is paralleled by glacial SST minimum. Maximum $\delta^{18}O$ value in core 17964-3 is also maximum glacial $\delta^{18}O$ value recorded in spliced record of both piston core (17964-2) and gravity core (17964-3) from this site (see Fig. 3c in Wang et al., 1999). C: $\delta^{18}O_{C.w}$ records of cores 17964-3 (open dots, solid line) and 17940-2 (solid dots, stippled line) versus age. Note that scale of core 17940-2 on right y-axis is shifted by 0.15‰ to align glacial $\delta^{18}O$ maxima of both cores. Vertical stippled lines are adapted from A and B. Age model of core 17964-3 (adapted from Pelejero et al., 1999a) gives age of onset of abrupt warming step of 14.6 ka, synchronous with two high-resolution records from neighboring cores 18252-3 and 18287-3 (Kienast et al., 2001b), whereas age model of core 17940-2 is based on radiocarbon dates in Wang et al. (1999). Note that benthic $\delta^{18}O$ record of core 17940-2 leads that of core 17964-3 at time of onset of Bølling warming by ~1400 yr. Only plausible way to bring benthic foraminiferal isotope records of both cores (recovered from comparable water depths) in phase during deglaciation is adopting increased ^{14}C reservoir age at northern site (17940-2) of ca. 1200–1300 yr. This approach, as well as magnitude of local increase in reservoir age at site 17940-2, is similar to record from North Atlantic of Waelbroeck et al. (2001). It further corroborates our assumption of synchronous deglacial warming throughout South China Sea (see text).

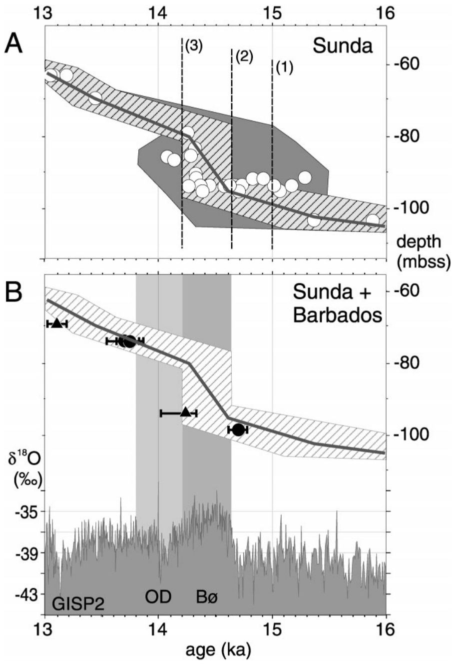

Figure 3. Estimated timing of meltwater pulse (MWP) 1a based on Sunda Shelf tidal-organic and Barbados coral-reef records as only data sets bounding pulse. A: Sunda Shelf sea-level reconstruction. Open circles represent age intercepts; note that numerous data points at 98–96 m are multiples produced by calibrating ^{14}C ages. Dark gray field: 1σ probability of calibrated ages; in parts strongly expanded due to prominent ^{14}C plateau. Vertical lines: (1) 15.0 ka: earliest possible start of MWP 1a given by 1σ probability of two dates (Hanebuth et al., 2000); (2) 14.67 ka: onset of MWP 1a if causatively linked (Hanebuth et al., 2000) to Bølling warming in Greenland (GISP2), and as evidenced by synchroneity of deglacial (Bølling) warming step and abrupt decrease in terrigenous organic matter supply to South China Sea (this study, see text); (3) 14.2 ka: end of MWP 1a based on last peak of uncalibrated ^{14}C ages 2σ probabilities (Hanebuth, 2000; Hanebuth et al., 2000). Hachured field: 1σ sea-level probability and assumed maximum paleotidal range (= vertical uncertainty) of ±3 m. Thick line is most probable sea-level curve (drawn by eye) after *n*-nonacosane concentrations from core 17940-2 (Fig. 1) and Hanebuth et al. (2000). B: Comparison of Sunda Shelf sea-level curve (from A) and Barbados data set (black triangles and circles) as summarized by Bard et al. (1998). Error bars indicate 2σ probabilities of single (filled triangle) and replicate (filled circles) analyses. Note that later start of MWP 1a (Fairbanks, 1989; Bard et al., 1990, 1996), which would coincide with Older Dryas (OD), is based on single U/Th date; omitting this sample, coral record would bracket redefined MWP 1a interval (see text). Bø—Bølling; mbss—meters below modern sea surface.

DISCUSSION

The first direct dating of MWP 1a from the South China Sea (Hanebuth et al., 2000) shows that it began between 15.0 and 14.4 ka and terminated no later than 14.2 ka (Fig. 3A). The termination of MWP 1a at 14.2 ka is delimited by the average of the youngest peak in the 2σ probability of the uncalibrated radiocarbon ages (Fig. 3A; Hanebuth, 2000). On the basis of an inferred coupling of the onset of MWP 1a and the Bølling warming in Greenland, Hanebuth et al. (2000) argued for an onset of MWP 1a at 14.6 ka (Fig. 3A). Our evidence for the close correspondence between the steep SST increase and the rapid drop in terrigenous sediment supply to the open South China Sea in sediment cores from the northern and southern South China Sea strongly implies that the Bølling warming and the onset of MWP 1a were synchronous; therefore the timing of MWP 1a is delimited as 14.6–14.3 ka.

There are several factors that need to be considered in attempting to reconcile the timing of MWP 1a based on coral records (Fairbanks, 1989; Bard et al., 1990, 1996) and the older age of MWP 1a presented here and by Hanebuth et al. (2000). First, corals (from monospecific reefs indicating a certain position with respect to the sea surface) can only yield maximum and minimum ages of rapid sea-level changes because they are drowned during sea-level rises faster than ~1.2 m/100 yr (Montaggioni et al., 1997). Thus, the bounds of the meltwater pulse are ultimately determined by the proximity of the coral samples to the central interval of most rapid rise in sea level (Fig. 3B). Second, there is a significant difference of 200–400 yr between the radiocarbon dates corrected for variable reservoir ages and changes in atmospheric ^{14}C (as far as known) and the U/Th ages of the corals used to define the timing of MWP 1a (Stuiver et al., 1998; Hanebuth et al., 2000). This offset is most likely due to uncertainties in the reconstruction of local reservoir ages, and thus usually only the U/Th dates are used for sea-level reconstructions. Consequently, the widely cited beginning of MWP 1a at 14.2 ka is defined by only a single U/Th date (from Barbados) without replicates (Fairbanks, 1989; Bard et al., 1990; Fig. 3B). Third, the onset of coral reef growth following land inundation is probably delayed by 200–300 yr due to the time needed for settlement and consolidation (Montaggioni et al., 1997). This delay would automatically reduce any closing age of the pulse determined from coral samples.

There are several implications of the phase relation between the Bølling warming and MWP 1a proposed here. First, the synchroneity of both events argues against any weakening of the thermohaline circulation in response to MWP 1a. To the contrary, according to the timing and phase relation proposed here, MWP 1a coincides with intensifying thermohaline circulation during the Bølling warming (Clark et al., 2002b). In turn, the absence of a slowdown in North Atlantic deep-water formation following MWP 1a could be taken as further evidence of an Antarctic source of the meltwater during MWP 1a (Clark et al., 2002a). Second, methane concentrations in ice cores (Severinghaus and Brook, 1999) show that the atmospheric content of methane increased rapidly during the Bølling warming. Given the apparent synchroneity of MWP 1a and the Bølling warming, we speculate that the methane increase could be partly due to an abrupt rise in water tables caused by heightened sea level. Third, given the immense impact of MWP 1a on coastal sedimentation, the proposed synchroneity cautions against the interpretation of changes in sedimentological parameters (e.g., grain-size distribution, organic carbon concentration) solely in terms of climatic and/or biological shifts that are presumed to have occurred synchronously with the Bølling warming. Finally, the synchroneity of MWP 1a and the Bølling warming in Greenland presented here is consistent with the general notion of rapid melting of ice sheets in response to dramatic warming events. According to a model study by Marshall and Clarke (1999), the large southward-extending vulnerable Laurentide ice sheet responded within a few years (i.e., synchronously within the resolution of marine sedimentary and coral records) to the air-temperature increase in excess of 5 °C at the onset of the Bølling period with a profusive meltwater pulse. According to Clark et al. (1996), however, there is little geological evidence for such a rapid melting of the Laurentide ice sheet. Furthermore, on the basis of a recent modeling experiment, Clark et al. (2002a) argued that most of the meltwater during MWP 1a must have originated from Antarctica. This could explain why there is no slowing of the thermohaline circulation in response to the massive freshwater input. The Antarctic origin of MWP 1a proposed by Clark et al. (2002a), however, poses the even greater conundrum of why there is no response (rapid ice-sheet melting) to a large forcing (Bølling warming) in the Northern Hemisphere, whereas there is an abrupt meltwater discharge from Antarctica, possibly without a similarly abrupt warming (Steig et al., 1998; but see Mulvaney et al., 2000; Grootes et al., 2001). In addition, glaciological studies so far indicate only a minor contribution of the Antarctic ice sheet to glacial sea-level lowering (e.g., Denton and Hughes, 2002).

CONCLUSIONS

A rapid drop in the supply of terrigenous organic matter to the open South China Sea also corresponds with a rapid increase in sea-surface temperature during the last deglaciation, corresponding with the Bølling warming at 14.7 ka. This is interpreted to reflect a rapid retrogression of local rivers due to rapid sea-level rise, strongly implying that the Bølling warming and the onset of MWP 1a are synchronous. This phase relation contrasts with the widely cited onset of this MWP 1a ca. 14 ka, and implies that MWP 1a did not cause a reduction of deep-water formation in the North Atlantic. To the contrary, following modeling studies by Mikolajewicz (1998) and Seidov et al. (2001), Clark et al. (2002c) proposed that an Antarctic source of MWP 1a (Clark et al., 2002a) could have actually caused the intensifying deep-water formation in the North Atlantic during this time, and the consequent Bølling warming. Our results demonstrating synchroneity of MWP 1A with the onset of the Bølling warming provide critical support for this hypothesis.

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