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Cooling of the South China Sea by the Toba eruption and correlation with other climate proxies ~71,000 years ago

Chi-Yue Huang

Department of Earth Sciences, National Chengkung University, Tainan, Taiwan

Meixun Zhao

Department of Earth Sciences, Dartmouth College, Hanover, NH03755, USA

Chia-Chun Wang

Department of Geology, National Taiwan University, Taipei, Taiwan

Ganjian Wei

Institute of Guangzhou Geochemistry, Academia Sinica, Guangzhou, China

Abstract. The Toba tephra layer has been identified in core MD972151 in the southern South China Sea (SCS), northeast of the Indonesia Toba caldera. This affords us an opportunity to directly determine a 1°C cooling for ca.1 kyr on the SCS following the Toba eruption (71 Ka) during the marine isotope stage 5a-4 transition, using century-scale sea surface temperature records. This cooling event in the SCS is well correlated with several coeval proxies such as increased East Asia winter monsoon intensity, increased ice-rafted detritus in the North Pacific Ocean sediments, and decreased $\delta^{18}O$ in Greenland ice core. Such correlation suggests a climate change where cold climate signals originated in the Northern Hemisphere ice sheets, transferred southward by the winter monsoon, and cooled the SCS.

Introduction

Modeling and climate proxy data have shown that large volcanic eruptions can spread gas and fine ash globally, and the resulted sulfuric acid in the stratosphere can cause significant global cooling [*Rampino and Self*, 1993; *Bekki et al.*, 1996; *Pollack et al.*, 1976]. For example, the Mountain Pinatubo eruption resulted in a global cooling of 0.5-0.7°C [*Dutton and Christy*, 1992]. Summer-temperature proxy records indicate that major volcanic eruptions have had significant influences on the major short term Northern Hemisphere coolings over the past 600 years [*Briffa et al.*, 1998]. The Toba eruption, the largest known late Quaternary explosive volcanic eruption dated to around 71-75 ka, could have resulted in an atmospheric loading to 1 x 10^{16} g of H_2SO_4 and produced a northern hemispheric cooling of 3-5°C for several years [*Rose and Chesner*, 1990; *Rampino and Self*, 1992]. The volcanic sulfate deposit of ~6-years in the GISP2 ice core about 71,000 years ago has been correlated with the Toba eruption and the associated $\delta^{18}O$ change corresponded to a cooling of 6-7°C [*Zielinski et al.*, 1996]. Such a dramatic chemistry and temperature change in atmosphere due to episodic volcanic eruption may also cause hazard for a decrease of human population [*Ambrose*, 1998; *Rampino and Ambrose*, 2000]. Few other high-resolution temperature records are available to assess the magnitude and duration of the Toba eruption on a global scale, especially in the ocean. Core MD972151 (8°43.73'N, 109°52.17'E, water depth 1,598 m, Figure 1), cored from the upper continental slope right below the Sunda Shelf of southern South China Sea (SCS), affords the opportunity to evaluate the cooling effect of Toba on the SCS region. Firstly, the identification of Toba tephra layer provides an unambiguous stratigraphic marker for the Toba tephra deposition in this core [*Song et al.*, 2000].

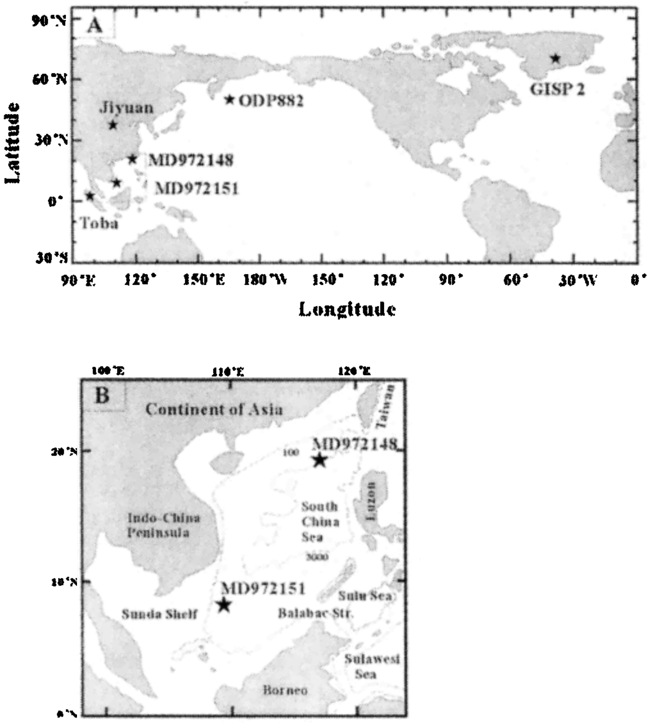

Figure 1. A map showing the locations of climate proxy records discussed in the text (A) and the detailed locations of two SCS cores studied in this paper (B). The Toba tephra layer is found in core MD972151 but not in core MD972148.

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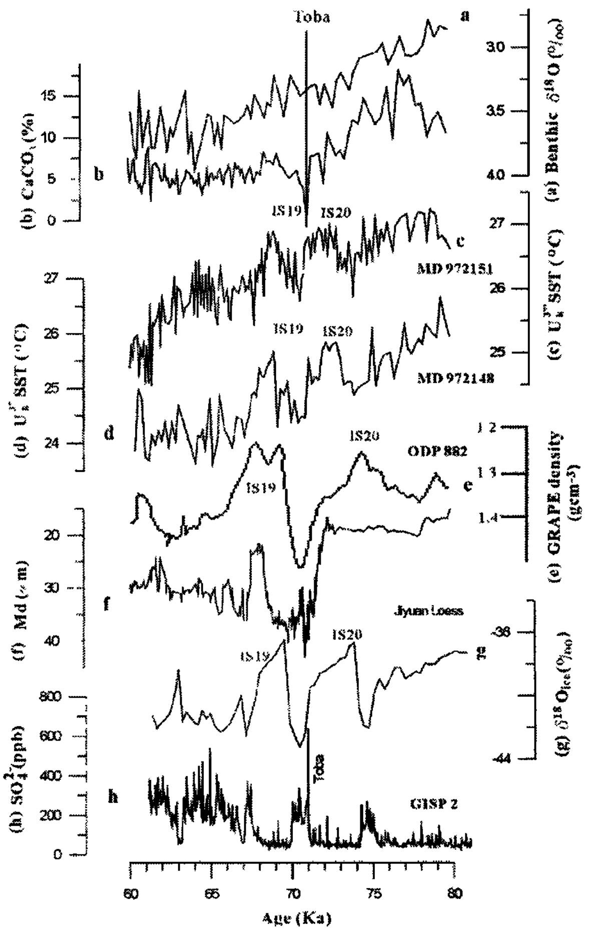

Figure 2. Climate proxy records for the period between 60,000 and 80,000 years ago. From top to the bottom: a) $\delta^{18}O$ (‰, PDB) record of benthic foraminifer *Cibicidoides wuellerstorfi* for core MD972151; b) Carbonate content (%, dry weight) for core MD972151, determined using a LECO WR-112 carbon analyzer; c) $U^{K'}_{37}$ SST (°C) for core MD972151. The method for alkenone analysis was described in *Zhao et al.* [1995] and SST was calculated using the calibration equation established using SCS core top sediments [*Pelejero and Grimalt,* 1997]: T (°C) = ($U^{K'}_{37}$-0.092)/0.031; d) $U^{K'}_{37}$ SST (°C) for core MD972148; e) GRAPE (Gamma Ray Attenuation Porosity Evaluator) density for ODP site 882 (g cm^{-3}). Data are from *Kotilaninen and Shackleton* [1995]; f) Loess mean grain size (μm) record for the JiYuan section located in the northern part of the Chinese Loess Plateau [*Ding et al.*, 1998]; g) $\delta^{18}O$ (‰, PDB) from the GISP2 ice core [*Dansgaard et al.*, 1993]; and h) Sulfate (ppb) from the GISP2 ice core [*Zielinski et al.*,1996].

The Toba label on the top indicates the sample containing the Toba tephra layer in SCS deep sea core MD972151. The Toba label on the bottom indicates Toba aerosols as identified by the sulfate peak in the GISP2 ice core.

Secondly, the high sedimentation rate of ca. 20 cm/kyr., due to high surface productivity [*Huang et al.*, 1997] and terrigenous input, allows us to reconstruct a century-scale SST record to assess the timing and duration of the SCS cooling in relation to the Toba tephra layer marker and to correlate this cooling event with other coeval climate proxy records in order to understand the mechanism of rapid climate changes and teleconnections.

Results

Stratigraphy for core MD972151 was established by AMS C-14 dating of 9 samples for the last 10,000 years [*Lee et al.*, 1999], and by correlating the benthic foraminifer *Cibicidoides wuellerstorfi* oxygen curve for this core with the SPECMAP $\delta^{18}O$ curve. The time scale of *Martinson et al.* [1987] is adapted. With this age model, the 2-cm Toba tephra layer has been assigned an age of ca. 71,000 years, during the marine isotope stage 5a to 4 transition (Figure 2). While this age is a few thousands years younger than the reported Toba eruption [*Rose and Chesner,* 1990], it is in agreement with the ice core chronology [*Zielinski et al.*, 1996].

Until recently, the Toba ash layer has been found mostly in the sediments of the Central Indian Ocean Basin, and has been used for regional stratigraphic correlation [*Rose and Chesner,* 1990; *Westgate et al.*, 1998]. A 2-cm layer at 1556-1558 cm (71ka) in the SCS core MD972151 with coarse glass shards in the southern South China Sea have been identified as tephra deposits from the Toba eruption [*Song et al.*, 2000]. Further evidence of its volcanic origin is provided by low magnetic susceptibility and the REE geochemistry of the glass shards [*Song et al.*, 2000], which show comparable characteristics to those of the Toba tephra [*Westgate et al.*, 1998]. The almost total depletion of carbonate provides further evidence for its volcanic origin (Figure 2b).

Long-chain ketones are produced by a few single-cell algae of the Class Haptophyceae. The unsaturation index [$U^{K'}_{37}$= (37:2)/(37:2+37:3)] of long-chain ketones in ocean sediments is positively related to annual mean growth temperature of common species, such as *Emiliania huxleyi* and *Gephyrocapsa oceanica* [*Prahl and Wakeham,* 1987]. In the SCS, $U^{K'}_{37}$ has been shown to reflect the annual mean SST for the surface layer (0-30 m) [*Pelejero and Grimalt,* 1997]. This index shows a glacial/interglacial cooling of ca. 4°C for the SCS [*Huang et al.*, 1997], but high-resolution records also show short-term changes in SST of 1-2°C over a few hundred years [*Huang et al.*, in preparation]. The MD972151 $U^{K'}_{37}$ SST record for the period 60,000 to 80,000 years ago (Figure 2c) also reveals several cooling events of ca. 1°C in several hundred years, one of them occurred immediately following the Toba eruption at ca. 71,000 years ago. The SST values for the few samples below and at the Toba tephra layer (73-71 Ka, corresponding to ice core IS 20) averaged 26.8°C. After the Toba eruption, SST then decreased to 25.8-26.0°C quickly and stayed generally low for ca. 1000 years but with a few oscillations (70.6-69.8 Ka; Figure 2c). Around 69 ka (corresponding to ice core IS19) SST returned to a comparable pre-eruption value of 26.8°C. The fact that this cooling occurred right after the Toba eruption may suggest that the Toba eruption was partially responsible for the SCS cooling around 71ka. In this case, other high-resolution SCS records would expect to also reveal this cooling event. Indeed, another high-resolution $U^{K'}_{37}$ SST record for core MD972148 (19°47.804'N, 117°32.56'E, water depth 2830m, Figure 1B) from the northern SCS also reveals a cooling of similar magnitude (1.2°C) and duration (1800 years) (Figure 2d). However, it should be pointed out that no tephra layer has been identified in core MD972148. Thus, the estimation of the timing and duration of this cooling event is based on an age model established using planktic and benthic foraminiferal $\delta^{18}O$ correlation, but not constrained by the tephra layer marker and other well-dated cores. Future high-resolution reconstruction in SCS is expected to identify the Toba tephra layer in more locations [*Bühring et al.*, 2000] and could further confirm this cooling event.

Correlation with other proxy records

Climatic evidence has suggested that the climates in the Asian continent/SCS and the northern high latitude can be linked via an atmospheric circulation mechanism [*An et al.*, 1990; *Ding et al.*, 1998]. Seasonally, SST in the SCS is basically controlled by the annually reversal of monsoons. During the winter season, SST is lower (18-20°C in the northern SCS and 26-27°C in the southern SCS) with a large north-south gradient. In contrast, SST is much higher and geographically homogenous (28-29°C) during the summer. Similarly, the SCS during glacial time was colder due to the strengthened East Asian winter monsoon [*Huang et al.*, 1997] which has been linked to the colder North Atlantic and polar climates via the westerlies [*Porter and An*, 1995; *Ding et al.*, 1998]. Furthermore, sub-Milankovitch scale cold events revealed by ice core $\delta^{18}O$ values are generally correlated with stronger Asian winter monsoon and lower SST in the SCS [*Huang et al.*, in preparation]. Such correlation provides further evidence pointing to a climate system where cold climate signals probably originated in the Northern Hemisphere ice sheets, transferred southward by the winter monsoon, and cooled the SCS.

Immediately following the Toba H_2SO_4 deposition, ice core $\delta^{18}O$ value decreased by 4.7‰ (Figures 2g and h), corresponding to a cooling of at least 6°C [*Zielinski et al.*, 1996]. More recently, this cooling event has been estimated to be as much as 16°C based on nitrogen isotope ratios of air bubbles trapped in GRIP ice core [*Lang et al.*, 1999]. The ice core cooling event lasted for about 1000 years (71-70 ka), almost the same time and duration of the SCS cooling (Figures 2c, d and g). If the cooling of the SCS and the Greenland region has been caused by the Toba eruption, then the fast response and teleconnection are consistent with a climate change involving an atmospheric circulation mechanism. In this case, the Toba cooling event could have been registered in other northern hemisphere high-resolution climatic records.

Grain size in the Chinese Loess Plateau is controlled by the intensity of the East-Asia winter monsoon, as grain size increases when winter monsoon strengths [*An et al.*, 1990]. Thus, for the Jiyuan section (Figure 1A), grain size is around 20 μm for the warmer and more humid interglacial period, but increased to 45 μm for the LGM when it was much colder and drier [*Ding et al.*, 1998]. Grain size in the Jiyuan section displayed a sudden and dramatic increase from a typical interglacial value of 20 μm (72 ka, IS20) to a glacial-like value of 40 μm in about 1000 years (Figure 2f), and stayed low for almost 3000 years. The grain size data signaled a sudden increase in the winter monsoon intensity at 71.8 ka, which must have cooled the east Asia continent and the SCS for thousands of years. This increase in the Asia winter monsoon occurred before the initiation of the stage 4 major glaciation. Rather, it was correlated with the cooling event between the warm interstadials 19 and 20 in the ice core records.

Ice-rafted detritus (IRD) in marine sediments reflects the discharges of icebergs to the oceans caused by climate changes. Sediments from the North Pacific Ocean are predominantly composed of low density biogenic opal and high density terrigenous materials [*Kotilaninen and Shackleton*, 1995]. Thus, variations in sediment density as measured by GRAPE (Gamma Ray Attenuation Porosity Evaluator) can be used to infer the relative contribution of IRD to the sediments. For the ODP site 882 record, the largest IRD peak during the last 80 kyr occurred around 72 ka and lasted for 1-2 ky [*Kotilaninen and Shackleton*, 1995](Figure 2e). This IRD peak has been correlated with the cold period between IS19 and 20 in the ice core record. Thus, the site 882 IRD event indicated a major cooling of the air temperature that controlled the area of glaciers entering the North Pacific, such as from the Okhotsk-Kamchatka region of Siberia, Aleutian Islands, southern Alaska, and the ice sheets north of the Bering Straits [*Kotilaninen and Shackleton*, 1995].

Three common features can be identified from the above climatic records regarding the cooling event occurred at ca. 71ka. (1), the cooling started following the eruption of Toba. For the SCS and ice core records, this timing was constrained by the identification of the Toba deposit layer. For the loess and North Pacific Ocean records, the timing was constrained by age models and correlation with the other records. We admit that there is an element of circularity in attributing the Toba eruption as the cause for the cooling event in these two records because their age models were partially derived from correlation with other well-dated records. However, such correlation at least suggests that the Toba eruption should be considered as a potential candidate for the trigger of this climate event. (2), the cooling lasted 1-3 kyrs according to the various age models. (3), although this cooling occurred during the marine isotope stage 5 to 4 transition, all proxies returned to values more typical of the warm periods (e.g., IS 19) following this cooling. Thus, these records clearly indicate that the Toba eruption did not cause the last major glaciation.

Discussion

The magnitude of this cooling registered in the SCS (1°C) and in the ice core (6 to 16°C) is consistent with modeling results for a volcanic eruption the size of Toba. Volcanic eruptions lead to different summer temperature cooling at different latitudes, with much larger decreases in higher latitudes but smaller decreases in low latitudes. For example, *Stothers* [1984] calculated that the Tambora eruption (1815 A.D.) caused a Northern Hemisphere temperature decrease of 0.7°C, but tree-ring data from the high latitude northern Quebec showed a summer temperature lowering of 3.5°C [*Jacoby et al.*, 1988]. GCM models showed that a hemispheric cooling could have been amplified at high latitudes by a factor of 4 to 7 [*Manabe and Bryan*, 1985; *Rampino and Self*, 1992]. Thus, a conservative estimate of an 3°C global cooling after the Toba eruption could have resulted in an 12°C reduction in the summer temperatures of the high latitude region. On the other hand, the low latitude region and especially the tropical ocean would experience less cooling. During the last glacial maximum (LGM), the tropical Pacific SST was lower by 1-2°C, while the high-latitude Atlantic SST was lower by 6-10°C, and ice core temperature was lower by 10°C [*CLIMAP*, 1976; *Thunell et al.*, 1994; *Dansgaard et al.*, 1993]. Thus, a 1°C SST decrease in the warm pool margin could correspond to ~5°C decrease in high latitude region, in agreement with magnitude of the Toba cooling registered in the SCS and GISP2 records.

However, the duration of this cooling event in the proxy records needs further consideration. The 6-year residence time of the aerosols suggests that the Toba eruption is unlikely to be directly responsible for the 1-3 kyr cooling in these records [*Zielinski et al.*, 1996]. However, the quantity and the longevity of the Toba aerosols in the stratosphere were sufficient to change the ocean-atmosphere system to achieve a steady-state temperature decrease, which was estimated to require 4 years of volcanic perturbations [*Pollack et al.*, 1976]. In such a system, the Toba ash cloud would increase the stratospheric aerosol albedo and reduce solar radiation receipts in the lower troposphere. The enhanced polar cooling and the subsequent growth of ice sheets would increase the Asia winter monsoon intensity, which in turn would cool China and transfer the cooling signal all the way to the tropical Pacific warm pool region. The cooling of the warm pool region would reduce the atmospheric water vapor content and further contribute to global cooling [*Bekki et al.*, 1996]. The establishment of this new but colder ocean-atmosphere climate system would certainly prolong the cooling initiated by the Toba eruption. It also worth noting that both the SCS SST records and the GISP2 record indicate that climates before the Toba eruption were experiencing major changes such as the cooling event before IS 20 at ca 75,000 years ago. Thus, the 71,000 year ago cooling event could as well have occurred without the Toba eruption. Toba may have just enhanced the magnitude and longevity of this cooling event. The lack of other oceanic records for the major cooling around 71,000 year ago may indicate that the Toba eruption did not cause a long-term ocean circulation change to cause a worldwide ocean cooling event. However, this may be a result of the available climate records lacking the high-resolution records covering the isotope stage 5.1 to 4 transition. More records and modeling studies would help to estimate the duration of such cooling. Evidence of cooling from other geographic region would also help to verify or disprove this mechanism.

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Chi-Yue Huang Department of Earth Sciences, National Chengkung University, 1 University Road, Tainan, Taiwan, ROC (e-mail huangcy@mail ucku.edu.tw)

Meixun Zhao, Department of Earth Sciences, Dartmouth College, Hanover, NH 03755, USA (e-mail: Meixun Zhao@Dartmouth.EDU)

Ganjian Wei, Institute of Guangzhou Geochemistry, Academia Sinica, Guangzhou 510640, China (e-mail: gjwei@gig.ac.cn)

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