Geo-Marine Letters
June 2008, Volume 28, Number 3, Pages 131-136
http://dx.doi.org/10.1007/s00367-007-0095-x

Evidence for the 8,200 a B. P. cooling event in the middle Okinawa Trough

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Abstract:

Based on new and existing data on oxygen isotopes, alkenone-surface seawater temperature trends, planktonic foraminifers, lithology, and clay mineral composition of piston cores, a distinct cooling event has been identified around 8,200 cal a B.P. in the middle Okinawa Trough, northwest Pacific. This corresponds to the 8,200 a B.P. cooling event recorded in many places of the Northern Hemisphere. During this event, the local temperature decreased by 1°C, and the $\delta^{18}O$ value increased by 0.6‰. A strengthened Asian winter monsoon is the most probable cause for this event, which thus adds further credibility to the contention that we are dealing here with a global phenomenon.

Introduction

Studies of abrupt climate changes are not only of academic interest, because of their influence on the evolution of life on earth, but in view of current global warming, also of social importance. Compared to the high-amplitude climate oscillations of the late stage of the last glaciation, paleoclimatic information recovered from ice cores, terrestrial lakes, speleothems, and marine sediments reveals that the Holocene climate appears to have been relatively stable. However, it is now widely recognized that rapid, small-magnitude climatic perturbations punctuated much of the Holocene for example, the Preboreal Oscillation, the 8,200 cal year B.P. cooling event (herein referred to as the 8,200 event), and the Little Ice Age (O'Brien et al. 1995; Bond et al. 1997). In particular, the 8,200 event has recently attracted interest worldwide, because it has left discernable signatures in many parts of the Northern Hemisphere, being generally characterized by cooler, drier, and perhaps more windy climate conditions (cf. review by Alley and Ágústsdóttir 2005). On the other hand, little evidence for this event has thus far been found in the Southern Hemisphere, the Pacific region, and East Asia. The East China Sea (ECS) is a typical open marginal sea in the northwestern Pacific. The sediments of the ECS record not only marine environmental changes, but also paleoclimatic changes on the Asian continent, especially those of the last glacial-interglacial cycle. The Okinawa Trough (OT) is a curved back-arc basin to the southeast of the ECS, washed by the Kuroshio Current (KC). During the Last Glacial Maximum (LGM), the sea level was about 135 m lower than today (Wang and Sun 1994), and most of the ECS continental shelf was thus emerged, the southern coastline bordering on the OT at that time.

Continuous, high sedimentary fluxes about 20 to 30 cm/ka to the trough (Xiong et al. 2005) favor high-resolution paleoclimatic and paleoceanographic studies in this region. Many authors have discussed different aspects of the period following the LGM, including millennial-scale climate fluctuations such as the Heinrich cooling events, the Younger Dryas (YD), or the paleohydrography of the region, e.g., the path variability of the KC in the late Quaternary (Shieh et al. 1997; Ujiié and Ujiié 1999; Jian et al. 2000; Li et al. 2001; Liu et al. 2001; Meng et al. 2002). However, to date no obvious 8,200 event signal has been reported in this case. This may be due to low sampling resolution, and decreasing sedimentation rates in the middle and northern OT, resulting from the landward migration of the Changjiang and Huanghe river mouths during the last transgression (Xiong et al. 2005).

In this paper, based on new and existing data on oxygen isotopes, alkenone-surface seawater temperature trends, planktonic foraminifers, lithology, and clay mineral composition of piston cores, we assess any evidence of the occurrence of the 8,200 cooling event in the northwest Pacific region.

Materials and methods

Piston cores DGKS9603 (28°08.869′N, 127°16.238′E) and DGKS9604 (28°16.64′N, 127°01.43′E) were recovered from the middle OT at water depths of 1,100 and 766 m, respectively, during the joint Chinese-French DONGHAI cruise in 1996 (Fig. 1). Except for two ash layers at 41–46 and 481–527 cm below seafloor, the 592-cm-long core DGKS9603 is composed mainly of gray silty clay (Li et al. 2001). Although Meng et al. (2002) have reported a cooling event around 7.0–7.8 cal ka B.P. in this core, they did not discuss this event in any detail, due to the interference of volcanic ash and glasses. Core DGKS9604, by contrast, is 1,076 cm long and composed mainly of clayey silt. In this case, the absence of volcanic interference, and the high sedimentation rate favored the search for the 8,200 event.

Fig1

High-resolution age models of both cores are based on the oxygen isotope record of *Globigerinoides sacculifer* ($\delta^{18}O$), and accelerator mass spectrometry (AMS) radiocarbon dating of planktonic foraminifers (Liu et al. 2001; Yu et al., unpublished data). The time resolution of successive samples from core DGKS9604 is around 100 years in the period following the LGM. For the reconstruction of sea surface temperatures (SST), in each case about 4 g of freeze-dried sediments was soaked with solvents and analyzed by gas chromatography for alkenone abundances. SST was calculated from the abundance of $C_{37:2}$ and $C_{37:3}$ alkenones using the formula of Müller et al. (1998):

$$\mathrm{SST} = (U_{37}^{k'} - 0.044) / 0.033$$

where modified ketone unsaturation index $U_{37}^{k'} = C_{37:2} / (C_{37:2} + C_{37:3})$.

For clay mineral analysis, samples from core DGKS9603 were treated with 15% H_2O_2 for 24 h to remove organic matter, then washed in distilled water, centrifuged, and finally dispersed in an ultrasonic bath. The fractions finer than 2 µm were separated according to Stokes' law, and mounted as texturally oriented aggregates by rapidly filtering the suspensions through membrane filters having 0.45 µm pore diameters. The filter cakes were dried under cool temperature, and mounted on aluminum tiles. They were solvated with ethylene-glycol vapor at 60 °C over 24 h before X-ray analyses. The measurements were conducted on a D/max-RB diffractometer with CuKα radiation (40 kV, 100 mA), the samples being X-rayed in the range 3–35° 2θ with a speed of 5 EG/min. Semi-quantitative evaluations of the mineral assemblages were made by applying the method of Biscaye (1965).

Results

Fig2,3

The 8,200 event manifests itself both in the $\delta^{18}O$ and the $U_{37}^{K'}$-SST signals in cores DGKS9603 and DGKS9604 (Fig. 2). The SST values decrease by about 1 °C, and the $\delta^{18}O$ values increase by 0.5 to 0.6‰. The amplitude is about 2/3 of that recorded in the YD event.

In addition, the foraminifer composition of core DGKS9603 changes substantially in the 43–40 cm layer (8.0–7.3 cal ka B.P.; Fig. 3). Thus, the contents of cool-water species such as *Neogloboquadrina pachyderma* (R) and *Globorotalia scitula* (Bé 1977) increase sharply in this core interval. At the LGM, the contents of *Globigerinita glutinata* and *Globigerina bulloides* are lower, whereas in the deglacial and postglacial periods the values increase, which probably indicates somewhat higher temperatures. In the 43–40 cm horizon, the contents of these two species decrease sharply, indicating a shift toward a cooler climate. The typical planktonic foraminiferal assemblage of core DGKS9603 thus reveals that around 8,200 cal a B.P., the SST in the middle OT decreased markedly. Although the abrupt shift in the δ^{18}O values in *G. sacculifer*, $U_{37}^{K'}$-SST, and typical planktonic foraminifer composition did not occur exactly at the same time, it is rather obvious that around 8,200 cal a B.P. a pronounced cooling event occurred in the middle OT.

In core DGKS9603, the 43–40 cm layer contains a substantial amount of volcanic glasses. As revealed in Fig. 4, the sediment concomitantly becomes coarser in this layer, consisting mainly of yellow silt and small amounts of clay. In addition, the carbonate content decreases, and the clay mineral composition changes sharply. The illite content decreases, whereas smectite, chlorite, and kaolinite increase.

Fig4

Discussion and conclusions

In the middle Okinawa Trough, the paleoceanography of the area was under the control of the Kuroshio Current, the Asian monsoon, and volcanic activity. It is difficult to determine which of these factors played the main role. In this context, the comparison of cores DGKS9604 and DGKS9603 has provided useful clues that helped to identify the leading factor responsible for the 8,200 event in the middle OT. Thus, smectite is generally derived from basic continental or marine volcanic materials, especially volcanic glasses. As it often occurs in areas with low sedimentation rates and near volcanic sources (Griffin et al. 1968), the high smectite content in the 43–40 cm layer suggests a strong influence of volcanic material. Based on the heavy mineral assemblage in the tephra, the origin of the smectite is probably associated with the Kikai-Akahoya (K-Ah) volcanic eruption in Japan (Okuno 2002). Chlorite, in turn, is a characteristic mineral of low-grade, chlorine-bearing metamorphic and basic source rocks, being very prone to chemical weathering. Its high content in the 43–40 cm layer indicates that some volcanic materials have been converted into chlorite, having probably been transported into the OT by wind. Illite, by contrast, tends to originate from more acidic terrestrial rocks, being relatively resistant to weathering (Biscaye 1965; Griffin et al. 1968). The decreased illite content in this layer suggests a decreased terrestrial input. Research on modern volcanic eruptions has shown that their influence on global temperature is relatively low. For example, the eruption of Mount St. Helens (Washington, USA) in 1980 lowered the air temperature by only 0.1 °C. Even the much larger eruption of the El Chichón volcano (Mexico) in 1982 resulted in a temperature decrease of only 0.3–0.5 °C. Mount Pinatubo in the Philippines injected 25 to 30 million tons of sulfur dioxide into the stratosphere during its eruption in 1991, lowering global temperatures by 0.5 °C (Tarbuck and Lutgens 2005). Even the largest volcanic eruption in the Quaternary, the Sumatra Toba eruption 74 ka B.P., resulted in a lowering of global SST by merely 1 °C (Oppenheimer 2002). These examples illustrate that the climate impact of a single volcanic eruption, no matter how large, is relatively small and short-lived. Although substantial quantities of volcanic glasses were found in core DGKS9603, there is no evidence of these in core DGKS9604, which was taken nearby. This indicates that a volcanic eruption did occur at the same time as the 8,200 event, but that it was small, and thus contributed little to the cooling event.

Fig5

The analysis of core DGKS9603 by Lv et al. (2002) showed that from 8.5 to ~8.0 cal ka B.P., the abundance of typical KC diatom species increased greatly, as did that of the main planktonic foraminifer indicator species of the KC, *Pullenium obliquiloculata* (Fig. 3). In addition, the trend of oxygen isotope composition of seawater ($\delta^{18}O_w$) reconstructed by Yu et al. (unpublished data) indicates that also the sea surface salinity increased at that time (Fig. 5). All of these observations suggest that the strength of the KC was enhanced during this period, which excludes it as a cause of the 'cooling' event.

Excluding volcanic eruptions, and changes in the KC as possible causes of the cooling event recorded in the OT, leaves a change in the activity of the winter monsoon as the only plausible alternative. Some authors have proposed that the cause of the 8,200 event around the North

Atlantic Ocean was associated with the final retreat of the Laurentide Icesheet during the deglacial period (e.g., Barber et al. 1999). Outbursts of glacial Lake Agassiz and Ojibway supplied large volumes of fresh water into Labrador Bay through Hudson Bay and the Hudson Strait, resulting in a slowdown of the Atlantic thermohaline circulation (Clarke et al. 2004). Due to the resulting change in the relative proportions of land, water and glacial ice in North America, this event was also recorded in Elk Lake in northwestern Minnesota (Walter et al. 2002).

Other studies have concluded that the winter monsoon was strengthened around 8,200 a B.P., as a result of which air temperatures were lowered over the Northern Hemisphere. Thus, Liu et al. (2003) found a peak in eolian yellow silt deposition around 8.4–8.2 ka B.P. in lake sediments of northwestern China, indicating an abrupt climate change associated with wind-dust storm events at that time. Wang et al. (1999) revealed that from 9.1 to 7.4 ka B.P., the climate of northern China was cold and dry. Wang et al. (2002) determined a temperature decrease of 7.8–10 °C around 9–8 ka B.P. from the Guliya ice core data. All these findings suggest that from 9 to 8 ka B.P., the strength of the winter monsoon was enhanced.

In the OT region, C_{37} alkenones are derived mainly from *Emiliania huxleyi* and *Gephyrocapsa oceanica*, which generally live within the euphotic zone in the upper 30 m of the water column (Tanaka 2003). Locally, these foraminifer indicator species, which reveal the 8,200 event, live within 25 m of the sea surface where maximum chlorophyll concentrations occur (Xu et al. 1999). In addition, the KC indicator species, *P. obliquiloculata*, lives mainly at water depths of up to 50 m.

Based on the above evidence, we suggest that the 8,200 event in the Okinawa Trough was manifested by a cooling of the upper 30–40 m of the local sea surface water, although the general hydrographic environment likely did not change dramatically. This implies that the East Asian monsoon played a major role in 'transporting' the cooling event signal from the North Atlantic Ocean to the Chinese continent and its marginal seas.

Acknowledgements

We would like to thank the crew of the DONGHAI cruise of the R/V *l'Atalante* in the East China Sea in 1996 for their support in sampling cores DGKS9603 and DGKS9604. We are indebted to Feng Guo for carrying out the clay mineral analyses of core DGKS9603. Reviews by Burg Flemming and two anonymous referees substantially improved this manuscript. This study was funded by the NSFC (grant no. 40421150011; 40431002), MOST (grant no. 2003cb716706), and SOA, China, as well as IFREMER, France.

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Fig. 1 Geographic locations of cores DGKS9603 and DGKS9604

Fig. 2 Variation of oxygen isotope $\delta^{18}O$ (‰, data from Liu et al. 2001) and $U_{37}^{K'}$-SST records (°C) in cores DGKS9603 and DGKS9604 (modified after Yu et al., unpublished data), compared with that of the Greenland GISP2 ice core (Stuiver et al. 1995). The *shaded lines* represent the 8,200 and YD events

Fig. 3 Typical foraminifer assemblage of core DGKS9603 over the last 20,000 cal years. The planktonic foraminifer data are from Li et al. (2001). The *gray shading* represents the layer in which foraminifer contents change sharply, coinciding with the 8,200 event

Fig. 4 Grain size, carbonate content (data from Liu et al. 1999), and clay mineral composition of core DGKS9603 over the last 20,000 cal years. The *gray shading*, which represents the 8,200 event, coincides with the layer in which these index parameters show marked departures from the general trend

Fig. 5 The $\delta^{18}O_w$ (Standard Mean Ocean Water) variation over the last 20,000 cal years in cores DGKS9603 (*gray curve*) and DGKD9604 (*black curve*; modified after Yu et al., unpublished data)

Fig. 1

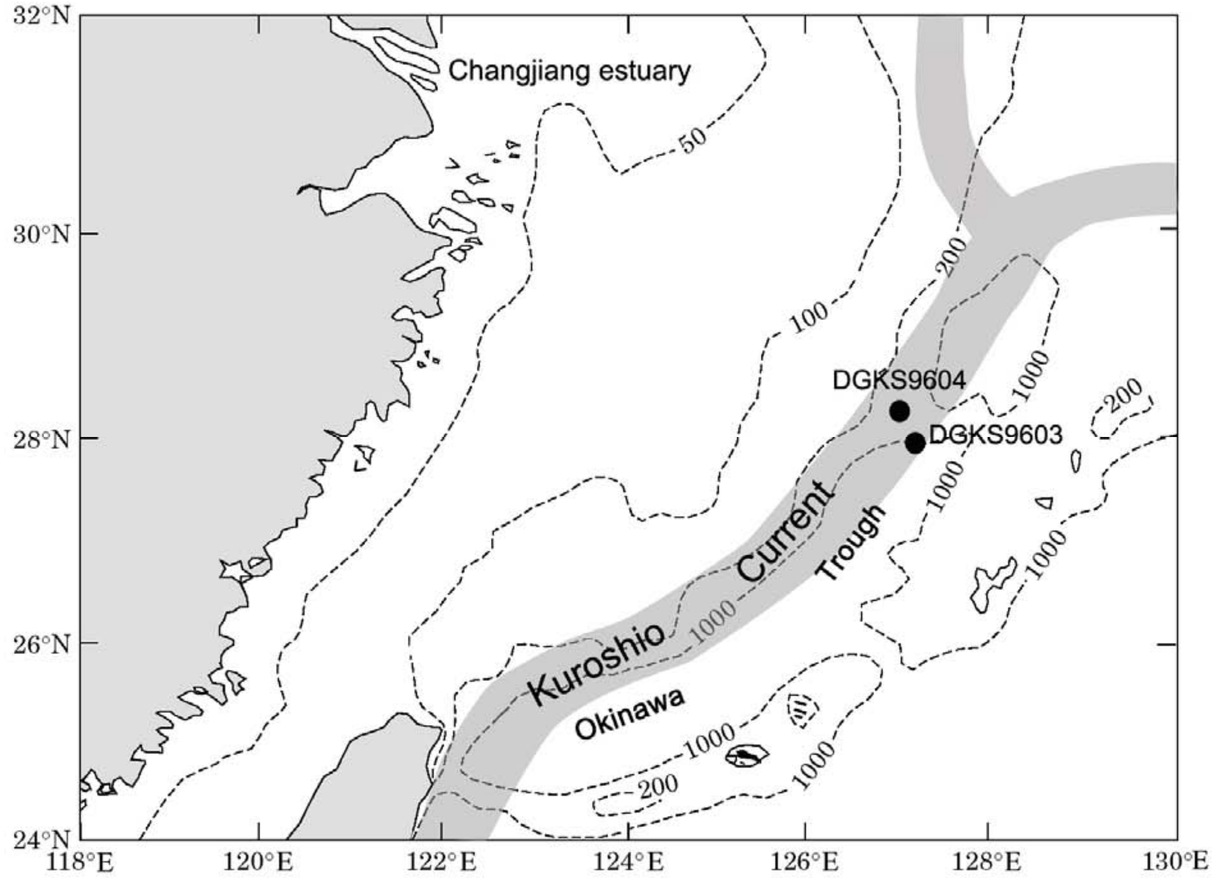
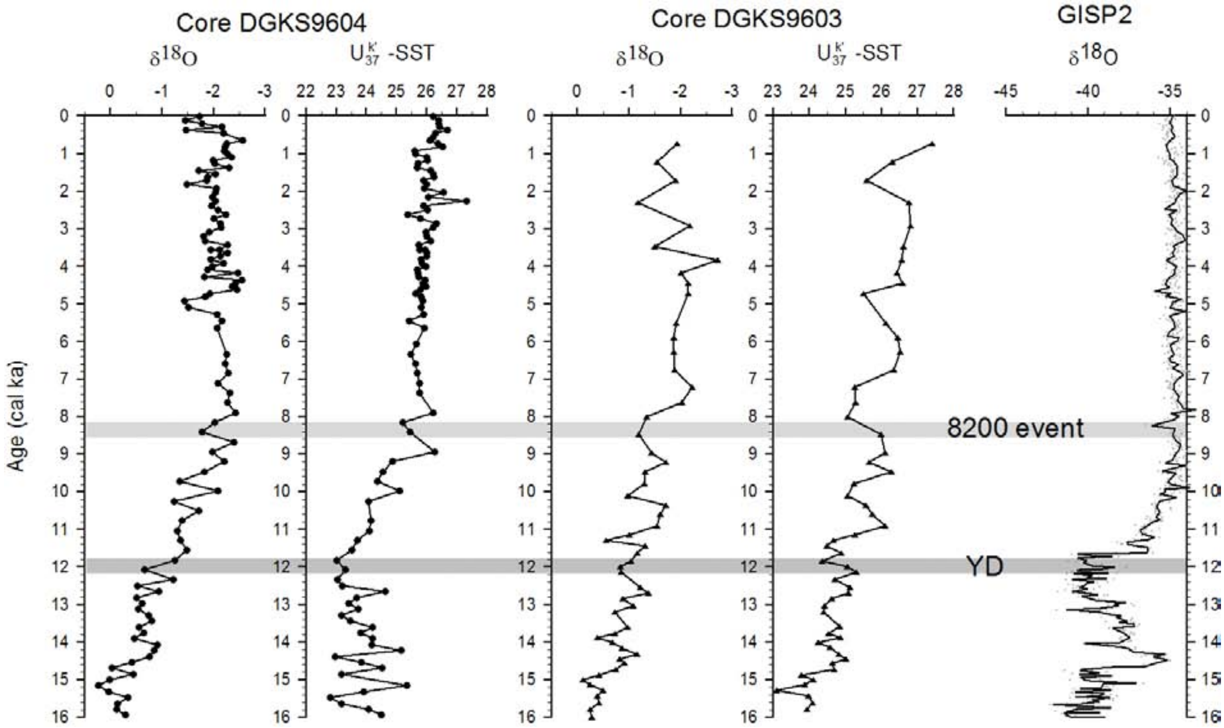
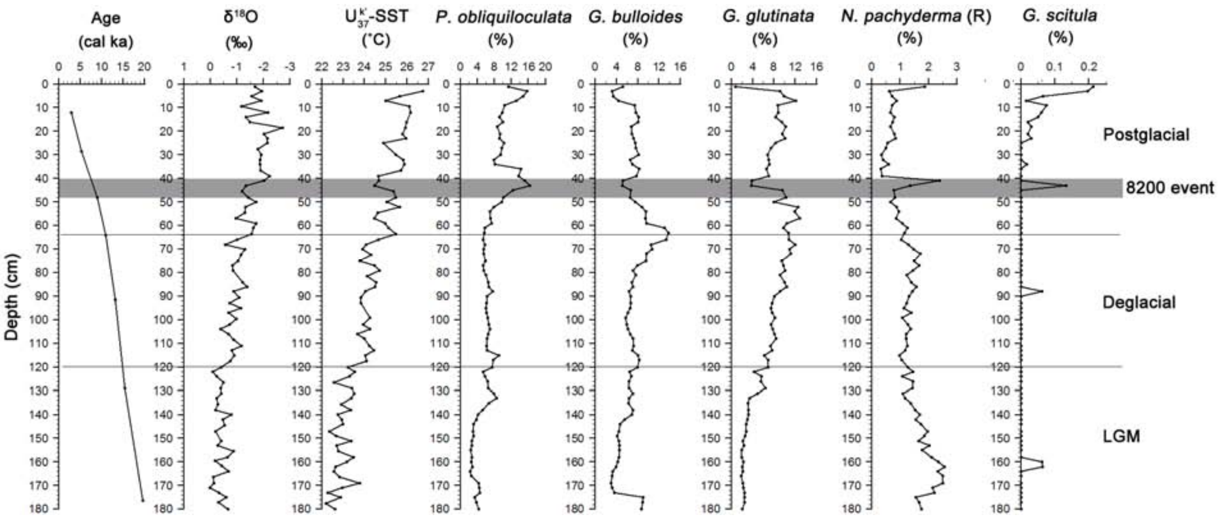

32°N
30°N
28°N
26°N
24°N
118°E
120°E
122°E
124°E
126°E
128°E
130°E
Changjiang estuary
50
100
200
1000
DGKS9604
DGKS9603
Kuroshio
Current
Trough
Okinawa

Fig. 2



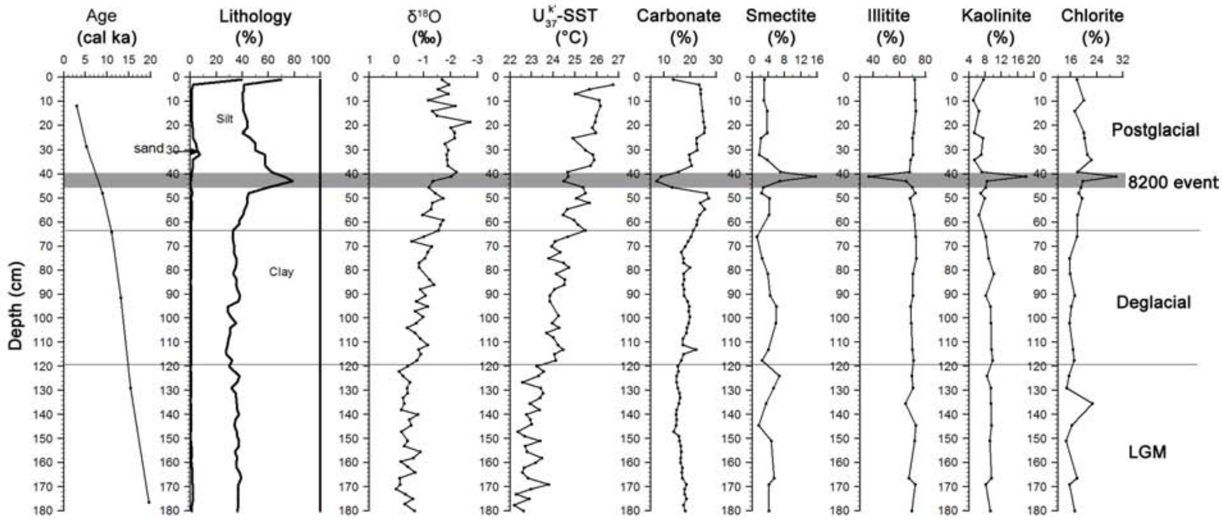
Core DGKS9604
Core DGKS9603
GISP2
$\delta^{18}O$
$U^{K'}_{37}$ -SST
Age (cal ka)
8200 event
YD

Fig. 3



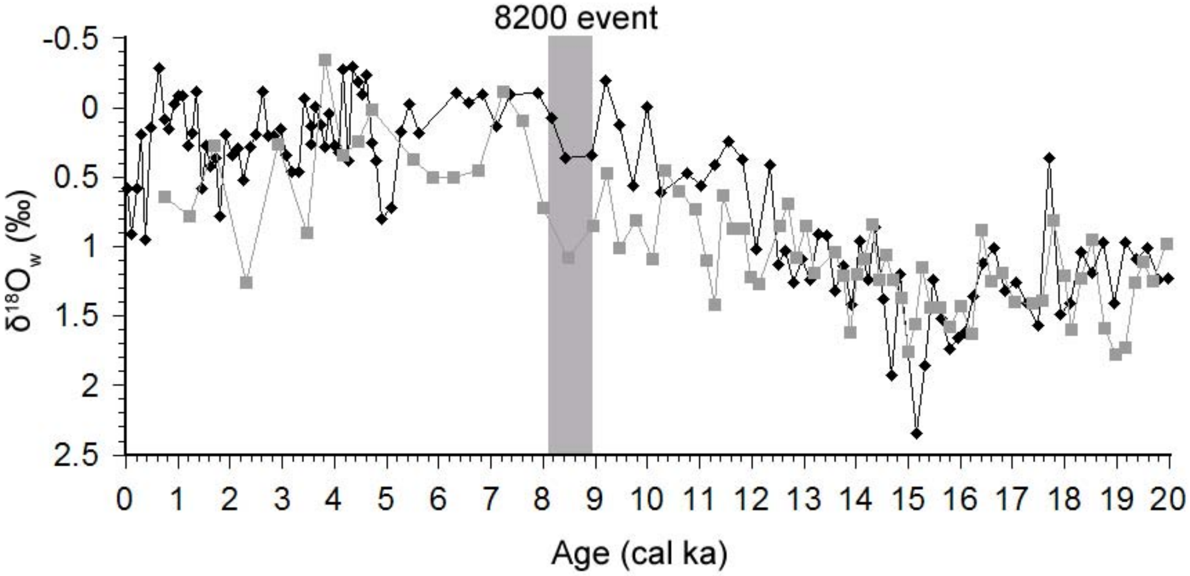
Age (cal ka)
δ18O (‰)
$U^{K'}_{37}$-SST (°C)
P. obliquiloculata (%)
G. bulloides (%)
G. glutinata (%)
N. pachyderma (R) (%)
G. scitula (%)
Depth (cm)
Postglacial
8200 event
Deglacial
LGM

Fig. 4



Age (cal ka)
Lithology (%)
δ18O (‰)
$U^{k'}_{37}$-SST (°C)
Carbonate (%)
Smectite (%)
Illitite (%)
Kaolinite (%)
Chlorite (%)
Depth (cm)
sand
Silt
Clay
Postglacial
8200 event
Deglacial
LGM

Fig. 5


8200 event
-0.5
0
0.5
1
1.5
2
2.5
δ18Ow (‰)
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20
Age (cal ka)