The Younger Dryas in the West Pacific marginal seas*

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Abstract The occurrence and nature of the Younger Dryas (YD) abrupt climatic event in the West Pacific marginal seas are discussed on the basis of 15 sediment cores. This event has been found in all these cores studied with a high-resolution stratigraphy and proved to be common to the West Pacific region. As shown by the isotopic and micropaleontologic analyses, the YD, dated by C-14 at about 11 000 to 10 000 a B.P., is a brief event of sea surface cooling in winter season following a fresh-water pulse about 12 000 a B.P. The "apparent regression" of the YD recorded in the Changjiang River delta and the Sea of Japan agrees with the interpretation that the YD is a period of slowed sea level rising between two phases of rapid rising. Both the winter surface water cooling and the increasing salinity in the YD imply a strengthening of the winter, but not summer monsoon circulation. This major climatic event in the marginal seas must have had profound impact on the adjacent continent.

Keywords: Younger Dryas, West Pacific, marginal sea, deglaciation, paleoceanography.

Our knowledge about the Younger Dryas, the most significant climatic reversal event within the glacial-Holocene transition, has progressed significantly in the recent years. As revealed by the Greenland ice cores, the Younger Dryas (YD) lasted 1 150 to 1 300 a and terminated abruptly over a period of approximately a decade[1]. The most common C-14 ages for the YD were 11—10 ka B.P., but recent studies show that the age is not precise mainly due to the long radiocarbon age plateau at 10 ka B.P. (Hajdas, 1994). Along with the discrepancy in age determination, two questions have been debated for years: Was the YD a northern Atlantic event only or a global phenomenon? What was the mechanism responsible for the event?

The West Pacific marginal seas, with their high sedimentation rates and enhanced glacial/interglacial contrast in climate, offer good opportunities to study the YD event. The extensive paleoceanographic investigations on these seas have resulted in numerous sites with stratigraphic resolution high enough to recognize the YD. The goal of the present paper is to review the published and our new data, in an attempt to show the nature of the YD in

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the West Pacific region.

1 Occurrences of YD in the region

Up to now, high-resolution sediment records have been obtained at least from 15 sites[1)] in the West Pacific marginal seas, and all have shown the YD event (table 1, fig. 1). In many of the cores the YD has been recognized from the $\delta^{18}O$ curves of shallow-dwelling species of planktonic foraminifers, while in others the event is revealed by the faunal composition or the transfer function paleotemperature estimations based on faunal composition of planktonic foraminifers.

Table 1 Sediment cores from the West Pacific marginal seas with records of YD event

Sea	Core	Latitude	Longitude	Water depth	Reference
Sea of Japan	KH-79-3	37° 4′ N	134° 42′ E	935 m	[2]
East China Sea	255	25° 12′ N	123° 6′ E	1 575 m	Li *et al.*[2)]
	253	25° 34′ N	123° 6′ E	839 m	Li *et al.*[2)]
	170	26° 38′ N	125° 48′ E	1 470 m	Li *et al.*[2)]
South China Sea	V36-3	19° 1′ N	116° 6′ E	2 809 m	[7]
	17 940	20° 7′ N	117° 23′ E	1 727 m	this paper
	SCS 90-36	17° 59′ N	111° 29′ E	2 050 m	Huang *et al.*[3)]
	SO 50-31KL	18° 45′ N	115° 52′ E	3 360 m	Huang *et al.*[4)]
	17962	7° 11′ N	112° 5′ E	1 969 m	this paper
Sulu Sea	ODP769A	8° 47′ N	121° 18′ E	3 643 m	[9]
	SO 58-69KL	8° 49′ N	121° 36′ E	4 696 m	[10]
	SO 50-31KL	8° 10′ N	121° 38′ E	4 911 m	[10]
Pacific off Japan	CH84-14	41° 44′ N	142° 33′ E	978 m	Kallel *et al.*, 1988
	C-1	36° 16′ N	141° 32′ E	1 545 m	[4]
	C-6	34° 43′ N	140° 33′ E	2 020 m	[4]

In the Sea of Japan, the last glacial maximum (LGM) was characterized by the development of thinly laminated clay layers indicative of stagnant bottom-water conditions caused by diluted surface water at the glacial lowered sea level stand, when the Sea of Japan was almost closed. In Core KH-79-3 (37° 4′ N, 134° 42′ E, w.d. 935 m, see fig. 1) a layer of thinly laminated clay reappeared and radiocarbon-dated at 10 800 a B. P., indicating a return to the stagnant conditions. Upwards it is followed by a positive shift of oxygen isotope value of planktonic foraminifers by 0.87 ‰ (fig. 2)[2]. This event is well comparable with the YD and may be interpreted as a reversal in mid-deglaciation related to a pause or even retreat in the post-glacial sea-level rising which has been reported from the

1) Three of the 15 sites are located outside the marginal seas, off the eastern coast of Japan.

2) Li, B. H., Jian, Z. M., Wang, P. X., *Pulleniatina obliquiloculata* as paleoceanographic indicator in the southern Okinawa Trough since the last 20 000 years, *marine Micropaleontology*, in the press.

3) Huang, C. Y., Wu, S. F., Zhao, M. X. *et al.*, Surface ocean and monsoon climate variability in the South China Sea since last glaciation, *Marine Micropaleontology*, in the press.

4) Huang, C. Y., Liu, P. M., Zhao, M. X. *et al.*, Marine and lake sediment records of East Asian monsoon, MS.

coastal areas of Japan[3]. Off the Pacific coast of Japan, isotopic and micropaleontological analyses of sediment cores have recorded a brief cooling event between 11 and 10 ka B.P. (Cores CH84-14, Kallel *et al.*, 1988; C-1 and C-6, see table 1 and fig. 1), and the authors believe that this was related to a southward shift of the polar front and readvance of the cold Oyashio Current at the YD[4].

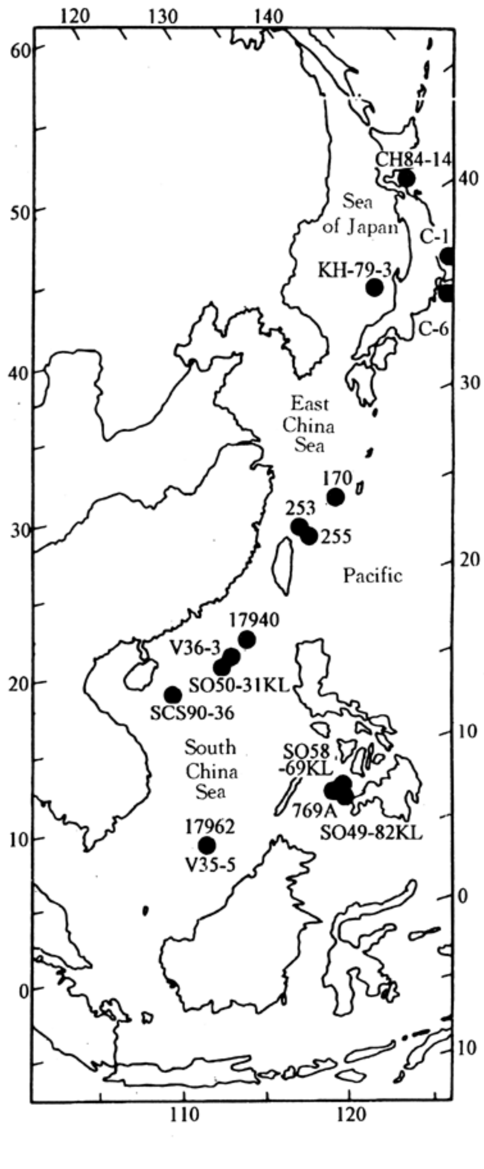

Fig. 1. Locations of the 15 sediment cores taken from the West Pacific marginal seas where the YD event has been recorded (see table 1).

The Okinawa Trough is the only area in the East China Sea where continuous sediment records through the glacial/Holocene transition are available. Core 255 (25° 12′ N, 123° 6′ E, w. d. 1 575 m) taken from the southern Okinawa Trough provides high-resolution section without turbidities. The YD event is evident from the percentages of warm-water species of planktonic foraminifers such as *Pulleniatina obliquiloculata* (fig. 3(d)) and from the oxygen-isotopic curve of *Globigerinoides sacculifer* (fig. 3(a)), but cannot be seen from that of benthic foraminifers (fig. 3 (b)). Similar patterns in downcore variations of *P. obliquiloculata* percentages can be seen from other cores from the southern (Core 253) and middle part of the Trough (Core 170; for locations see table 1 and figure 1)[1)].

Much more sediment cores have been studied from the South China Sea (SCS), although the occurrence of YD there was questioned. Broecker *et al.* (1988)[5] studied Core V35-5 from the southern SCS (7 ° 12′ N, 112° 5′ E, w.d. 1 953 m, fig. 1) with numerous AMS C-14 datings and found an abrupt termination of the last glacial period, which was used as an evidence for the absence of the YD in the Pacific. However, the abundances of two warm-water planktonic foraminifers they used, *G. sacculifer* and *P. obliquiloculata*, were based on weight not on number of specimens, and even so the glacial-Holocene transit was not straight-lined (fig. 4). During the German-Chinese joint paleoceanographic cruise in

1) See footnote 2) on page 523.

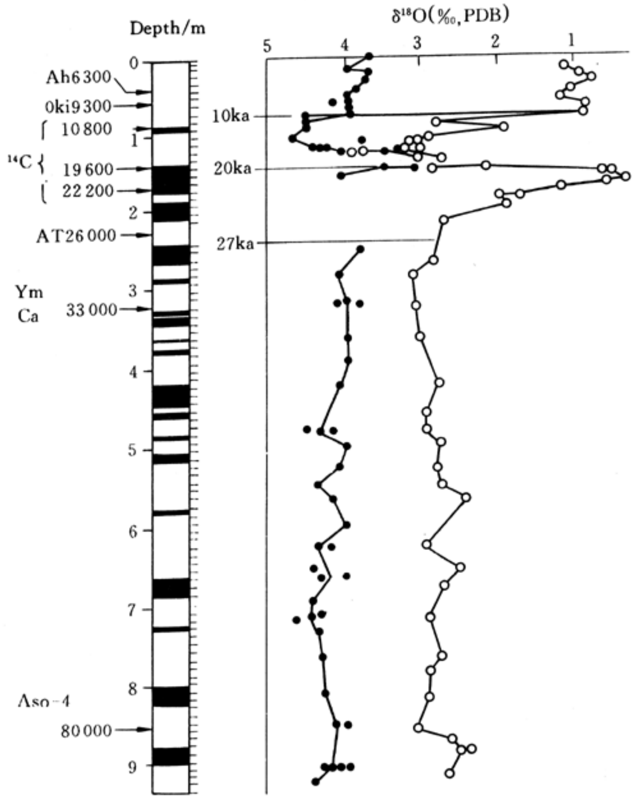

Fig. 2. Oxygen-isotope variations in Core KH-79-3 from the Sea of Japan. Black sections in column indicate horizons of thinly laminated clay; ●, benthic foraminifers; ○, planktonic foraminifers. Ages shown on the left are based on tephrochronology (Ah, Oki, AT, Ym and Aso) or radiocarbon datings. For location see fig. 1 (Redrawn from ref. [2]).

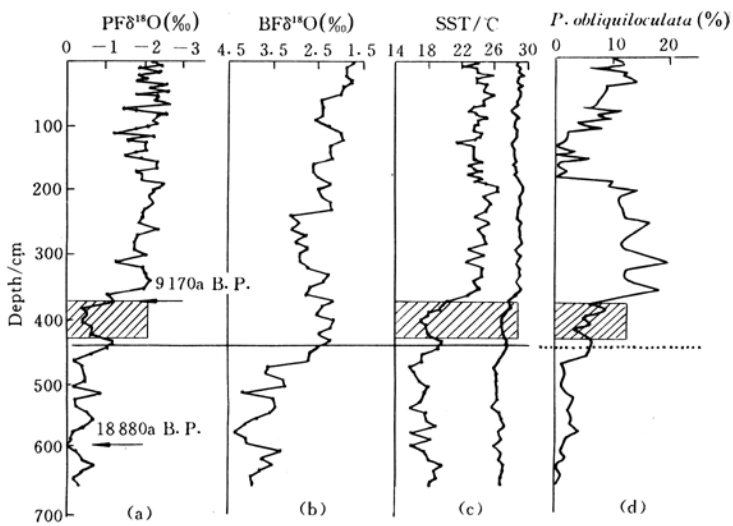

Fig. 3. Isotopic and micropaleontologic data of Core 255 from Okinawa Trough. (From Li *et al.*[1] for location see fig. 1). (a) Oxygen isotope curve of planktonic foraminifer *G. sacculifer*; (b) oxygen isotope curve of benthic foraminifer *Cibicides wuellerstorfi*; (c) paleotemperature curve (left: winter, right: summer) based on estimates using Transfer Function FP12-E; (d) percentage of *P. obliquiloculata* in planktonic foraminiferal fauna, indicative of winter SST.

1) See footnote 2) on page 523.

the SCS[6] we have recorded the same site, and Core 17 962 (7° 11′ N, 112° 5′ E, w.d. 1 969 m, fig. 1) shows the same stratigraphy. Our preliminary micropaleontological analysis has revealed that the YD reduction in the warm-water species abundances is well comparable to that in other cores, although the magnitude of decrease is less significant (fig. 4).

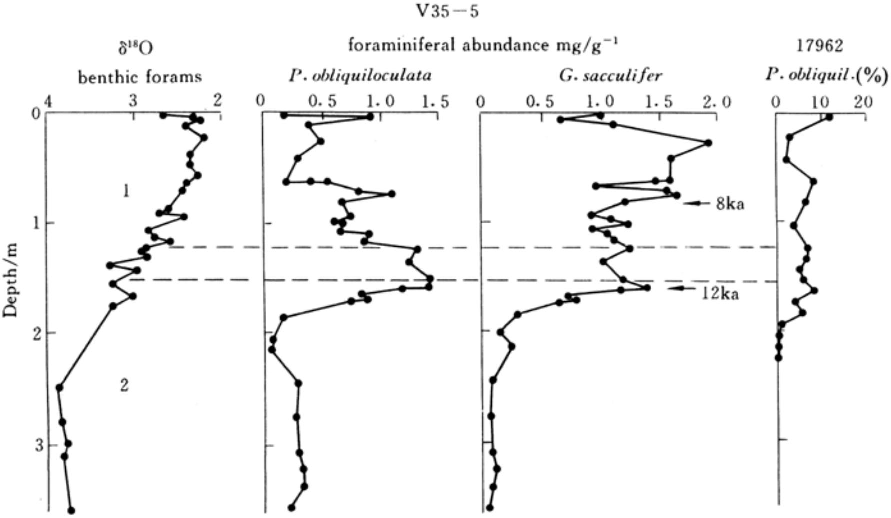

Fig. 4. Downcore variations of warm-water planktonic foraminifers in Core V35-5 and Core 17962 taken from the same location in the southern South China Sea (see fig. 1). Dotted lines indicate the YD event with a cooling signal less prominent than in the northern part of the sea. The foraminifer abundances in Core V35-5 are based on weight (milligram per gram of sediment), whereas percentage of *P. obliquiloculata* in Core 17962 is based on number of specimens (V35-5 data from reference [5]).

From the northern continental slope of the SCS, Core V36-3 (19° 1′ N, 116° 6′ E, w.d. 2 809 m) displays a clear reversal event corresponding to the YD both in the oxygen isotopic records of *G. sacculifer* and in downcore abundance variations of warm-water species such as *P. obliquiloculata*[7, 8]. The YD event has been recorded both isotopically and micropaleontologically in other cores from the northern slope (fig. 1, table 1): Core 17 940 (20° 7′ N, 117° 23′ E, w.d. 1 727 m) taken by the German-Chinese joint cruise[6], Core SCS 90-36 (17° 59′ N, 111° 29′ E, w.d. 2 050 m)[1)], and Core SO 50-31 KL (18° 45′ N, 115° 52′ E, w.d. 3 360 m) where the YD has been dated at 11 390—10 310 a B.P.[2)]

From the Sulu Sea high-resolution stratigraphy with records of the YD event has been reported from at least 3 cores (fig. 1, table 1): the uppermost part of ODP 769A (8° 47′ N, 121° 18′ E, w.d. 3 643 m)[9], Cores SO 58-69 KL (8° 49′ N, 121° 36′ E, w.d. 4 696 m) and SO 49-82 KL (8° 10′ N, 121° 38′ E, w.d. 4 911 m)[10]. The YD is clearly marked by the increase of $\delta^{18}O$ value of *Globigerinoides ruber* and the reappearance of cool-water planktonic foraminifers (figure 5)[9].

1) See footnote 3) on page 523.

2) See footnote 4) on page 523.

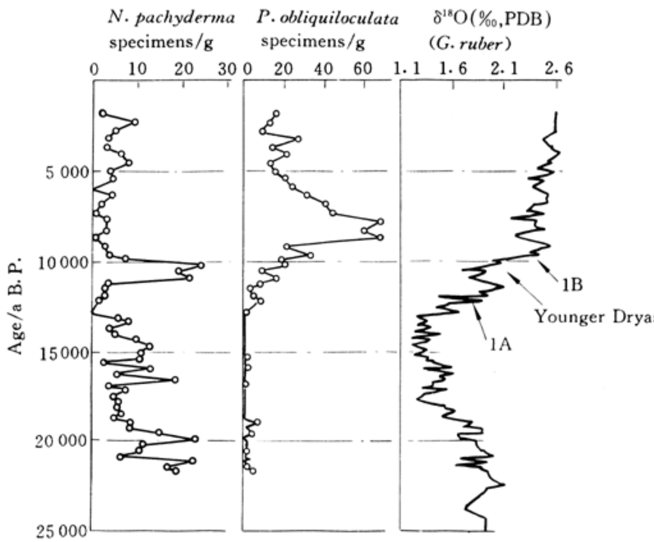

Fig. 5. Isotopic and micropaleontological data of ODP Core 769A taken from the Sulu Sea. 1A and 1B are two phases of rapid sea-level rising before and after the YD event. Notice the high absolute abundance of *N. pachyderma* (dextral) in the YD comparable with that in the LGM and its lower relative abundance in the YD, as the abundance of warm-water species such as *P. obliquiloculata* was much higher in the YD than in the LGM. For location see fig. 1 (redrawn from reference [9]).

2 Characteristics of YD

In the North Atlantic, the YD is recognized as a major cooling and a southward migration of the polar front, separating the two steps of the deglacial warming (Termination 1A and Termination 1B). What are the characteristics of the YD event in the West Pacific marginal seas? Faunal and isotopic analyses of the 15 cores (fig. 1, table 1) have enabled us to address the question from three aspects: sea surface temperature (SST), salinity and sea-level changes.

2.1 SST

As a basic indicator of the YD, the sea surface water cooling has been noted in all the sites with the YD records. Not yet clear is the magnitude of the temperature decline. Its estimations range from a negligible oscillation for some authors[11] to a return to sea surface conditions "comparable to those of the glacial maximum"[12]. Recent extensive paleoceanographic studies in the region have made possible a closer view of the YD SST conditions.

The most paleo-SST reconstructions in the West Pacific region are based on planktonic foraminifers using Transfer Function FP12-E (Thompson, 1981). The studies of the cores from the China Seas show that the Glacial-Postglacial SST contrast has been recorded mainly for winter season, while the summer SST displays very little changes. For example, the glacial-Holocene SST contrast in the middle and southern Okinawa Trough has been estimated to be 6.8—9.9℃ for winter, and 2.0—3.5℃ for summer[1]. At the site V36-3 in the northern SCS, the contrast is 5.7℃ for winter and 2.1℃ for summer[7]. Thus, the glacial cooling deals mainly with the winter SST, while the difference in summer SST is minor, hardly exceeding the standard error for the Transfer Function FP12-E.

The same applies to the YD cooling which, again, are seen mainly from the winter SST curves. As shown by the estimations from five cores of the East China Sea and the northern SCS (table 2), the YD event was a 1.5—3.3℃ winter SST cooling in the midway of the deglaciation warming, and the SST did not return to the full glacial conditions. The YD minimum SST for winter was still 1.5—3.5℃ higher than that of the LGM. Let it be noted that the transfer function SST estimation has been challenged by other paleotemperature techniques, in particular for the tropics[13]. For example, long-chain alkenones from algae (U_k^{37}) provide an alternative for paleo-SST estimations. Table 2 shows some results of SST estimations for two cores using U_k^{37}, where the YD drop at SST is estimated to be only 0.5℃ and the YD-LGM difference at winter SST only 1—1.3℃, significantly smaller than the transfer function SST estimations. The discrepancy is at least partly explained by the fact that the U_k^{37} paleo-SST estimations do not distinguish winter values from summer values.

Table 2 SST variations between the LGM, DR and Holocene optimum in the West Pacific marginal seas

Site		SST		Data source
sea	core	YD retreat[a)]	YD-LGM difference	
winter SST based on transfer function FP-12E				
Off Japan	C-2	~2	2—4	see text
	C-6	~2	2—4	see text
East China Sea	255	2.5	1.5	see footnote 2) on page 523
	170	2	1.5	see footnote 2) on page 523
	253	3.3	3.5	see footnote 2) on page 523
South China Sea	V36-3	2.7	3	[7]
	17962	1.5	3	new data
SST based on U_K^{37}				
South China Sea	SCS90-36	0.5	1	see footnote 3) on page 523
	SO 50-31 KL	0.6	1.3	see footnote 4) on page 523

a) "YD retreat" is the difference between the YD minimum value and the value of the preceding warming representing Termination 1A.

In the Sea of Japan no high-resolution records of paleo-SST have been reported. Two cores off the eastern coast of Japan are used here to show the SST variations in the northern part of the West Pacific margin (table 2). If the linear relationship

1) See footnote 2) on page 523

between the modern SST and modern composition of plankton groups is applied (see fig. 11 of ref. [4]), the YD retreat of SST reaches only about 2℃ in SST estimations for Cores C-1 and C-6, and the difference between the YD minimum and Holocene optimum SST is about 4℃. Since the glacial/Holocene SST contrast is as high as 6—8℃ for this area (Moore *et al.*, 1980), the SST at the YD must have been 2—4℃ higher than at the LGM (table 2). Studying cores off the east coast of Japan, Kallel *et al.* (1988) and Chinzei *et al.* (1987)[4] both related the YD cooling to the southward readvance of the polar front.

2.2 Sea level

Another interesting question is whether the YD cooling had a sea-level change signal. Using radiocarbon-dated submerged coral reefs from Barbados, Fairbanks was able to show that the deglacial sea-level rise was not monotonic; rather, it was marked by two intervals of rapid rise centred at 12 000 and 9 500 C-14 years B.P. (or 13 070 and 10 445 calendar years B.P.), and the YD was a period of low rate of sea level rise in between[14]. Thus, the YD cooling, unlike the glacial, was not accompanied by ice-cap growth and its corresponding oxygen-isotopic signal. If the oxygen-isotopic curves of planktonic and benthic foraminifers are compared, the YD shift can be seen only from those of planktonic species, but not from the benthic (e.g. Cores SCS90-36 from the SCS; Core KH-79-3 from the Sea of Japan, fig. 2). The only exception is Core SO58-69KL where the YD is displayed also on the curve of *Bulimina* sp[10]. Although the finding requests further studies and explanation, it is believed that the YD did not cause a growth of the global ice-cap.

However, a sea-level drop signal of the YD was discovered on the Changjiang (Yangtze) River delta. Here, a thin foraminifers-bearing layer, with AMS C-14 dating at 11 000 a B.P., has been found in bore holes on the coast of the eastern end of the delta. This coastal-marine layer is overlaid by non-marine silts until the Holocene marine transgression deposits started, but it disappears landward in a couple of kilometers. We believe that the transgression/regression history of the Changjiang River delta is an expression of the dynamic balance between the eustatic sea-level change rate and the rate of sediment accumulation. The brief transgression before the YD corresponds to the first rapid sea-level rise phase according to Fairbanks[14], and the apparent brief regression of the YD age must have resulted from the surplus accumulation rate of sediments exceeding the rate of sea-level rise slowed down with the YD event. Since the Holocene transgression reached the China coast plains later, the YD signal can be found only locally.

Another sea-level signal occurs in the Sea of Japan. Thinly-layered clay marks the LGM there when the sea was almost closed due to the lower glacial sea-level, but this thinly-laminated clay reappeared at 10 800 a B.P. in Core KH-79-3[2] (fig. 2). Again, this might imply a return to certain isolation of the sea related to “apparent” regression caused by the YD interruption of global sea-level rise as mentioned by Japanese coastal studies[4].

2.3 Salinity

Did salinity of the sea surface water change in the West Pacific marginal seas at the YD cooling? This is another interesting and debatable question. Marginal seas are most sensitive to salinity changes because of its adjacency to the land, but it remains difficult to extract paleo-salinity information from isotopic data for the Pacific region, leaving room for debates and speculations. Based on the Barbados data, Fairbanks[14] found $\Delta\delta^{18}O$ between Holocene and LGM to be 1.3 ‰ for the global ice-cap and sea-level change, and the YD occurred when the oceanic water $\delta^{18}O$ was 0.6 ‰ heavier than now if a pure ice-volume effect is considered. Any difference from the global signal must be caused by local or regional factors such as SST or salinity. Kudrass *et al.*[10], for example, ascribed the 0.9 ‰ difference (between global and local) in $\delta^{18}O$ at the YD in Core SO49-82KL totally to SST, and estimated a 3℃ cooling in surface water for the Sulu Sea. On the other hand, Linsley and Thunell[9] believed that the 0.4 ‰ YD shift in $\delta^{18}O$ found in Core ODP 769A from the same Sulu Sea was a salinity signal, with the surface water salinity 1 ‰ lower before the YD. This idea well correlates with the deglaciation model[14] with two melt-water pulses at 12 000 and 9 500 a B.P. giving rise to phases of rapid sea-level rise bracketing the YD event.

A similar salinity change must have occurred in the SCS, at least in its northern part where high-resolution analyses have been conducted. An example is Core SCS90-36 where the YD retreat of 0.5 ‰ in $\delta^{18}O$ is revealed only from the curve of the surface-dwelling species *G. sacculifer*, not from that of deep-dwelling species, *Globorotalia menardii*, not to say the benthic species. This is a convincing evidence for salinity change: a strong fresh-water pulse around 12 000 a B.P. has led to a much lighter oxygen-isotope value in the northern part of the SCS, and the decrease of the fresh-water input at the YD event has caused a return to an increase of surface-water salinity and, hence, led to heavier oxygen-isotope. However, all these changes happened only within the surface water recorded only by shallow dwelling, not by deeper-dwelling species inhabiting the subsurface water.

The above-mentioned thinly-laminated clay deposited in the LGM and YD on the bottom of the Sea of Japan is, again, related to salinity changes. This clay was accumulated under strongly anoxic bottom conditions in a well stratified water column below diluted surface water, when the sea was somehow closed[2]. If this is correct, the surface water of the YD Japan Sea should be of lower salinity. Another salinity effect was proposed by Keigwin and Gorbarenko[11] for a site at the Tsugaru Strait, the connection of the Sea of Japan with the Pacific. The positive YD shift of $\delta^{18}O$ (again only in planktonic not in benthic curve) in Core CH84-14 was assigned to a decrease of SST by Kallel *et al.* (1988), but Keigwin and Gorbarenko[11] considered the lighter $\delta^{18}O$ value before the YD as a result of the abrupt sea-level rise at 12 000 a B.P. when diluted surface water from the Sea of Japan discharged to the open Pacific.

3 On origin of the YD event

Many publications have been devoted to the origin of the YD event, and its records in the West Pacific marginal seas may throw new light to its origin. Probably two opposite interpretations of the YD are the most important: According to Broecker *et al.*[15] the YD represents a brief return to glacial-like conditions triggered by melt-water discharge to the North Atlantic and, therefore, is a regional phenomenon restricted to the northern Atlantic. Fairbanks[15], on the contrary, considered the YD as a period of low rate of sea level rise between the two large melt-water pulses which are global in scale. Our West Pacific evidence for the YD, together with the above-described salinity and sea-level signals, clearly support the Fairbanks' hypothesis.

Another aspect is the possible change in atmospheric circulation related to the YD event. Since the East Asian monsoon system is the most remarkable climatic feature for the region in question, it is essential to know if the monsoon circulation was changed with the YD cooling. An *et al.*[12] suggested that the YD age was characterized by a strengthening of summer monsoon in China leading to the enhanced pedogenesis in the Loess Plateau. For explanation they suggested that YD oceanic conditions in the West Pacific were "comparable to those of the glacial maximum", and the cool ocean had given rise to intensified summer monsoon bringing moisture to the hinterland. The hypothesis proposed is, unfortunately, not supported by the marine records in the West Pacific marginal seas. As discussed above, the YD winter SST there was 1.5—3.5℃ higher than that in the LGM, and the summer SST had experienced only negligible variation at the YD. Therefore, there is no reason to suggest any significant cooling of SST in summer which could strengthen the summer monsoon. Rather, there is evidence to support a weakened summer monsoon and strengthened winter monsoon circulation for the region, including the reduced fresh-water input to the marginal seas and the noticeable decrease in winter, not summer SST at the YD time.

4 Conclusions

(1) The Younger Dryas event has been found at least in 15 sediment cores taken from the Sea of Japan, East China Sea, South China Sea, Sulu Sea etc. and is shown to be a common phenomenon in the West Pacific marginal seas.

(2) According to the estimations in the studied cores, the YD cooling in the West Pacific marginal seas deals mainly with winter SST with an SST retreat by 1.5—3.3℃ for winter, while the YD change in summer SST was very insignificant. The decreased winter SST was still about 1.5—4℃ higher in the YD than in the LGM.

(3) The "apparent" regression at the YD as recorded in the Changjiang River delta and in the Sea of Japan supports the interpretation of the YD event as a period of slowed

sea-level rise between two steps of rapid rising caused by two major melt-water pulses.

(4) There is salinity signal in the YD $\delta^{18}O$ shift recorded in the studied seas. The restriction of the $\delta^{18}O$ shift only to the shallow-dwelling species implies a salinity rise at the YD as compared to the preceding stage.

(5) The decrease of SST and the increase of salinity in surface water strongly suggest a strengthening of winter, but not summer monsoon circulation for the YD in the West Pacific marginal seas.

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