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Exploring cyclic changes of the ocean carbon reservoir

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Abstract **A 5-Ma record from ODP Site 1143 has revealed the long-term cycles of 400—500 ka in the carbon isotope variations. The periodicity is correlatable all over the global ocean and hence indicative of low-frequency changes in the ocean carbon reservoir. As the same periodicity is also found in carbonate and eolian dust records in the tropical ocean, it may have been caused by such low-latitude processes like monsoon. According to the Quaternary records from Site 1143 and elsewhere, major ice-sheet expansion and major transition in glacial cyclicity (such as the Mid-Brunhes Event and the Mid-Pleistocene Revolution) were all preceded by reorganization in the ocean carbon reservoir expressed as an episode of carbon isotope maximum (δ^{13}Cmax), implying the role of carbon cycling in modulating the glacial periodicity. The Quaternary glacial cycles, therefore, should no more be ascribed to the physical response to insolation changes at the Northern Hemisphere high latitudes alone; rather, they have been driven by the "double forcing", a combination of processes at both high and low latitudes, and of processes in both physical (ice-sheet) and biogeochemical (carbon cycling) realms. As the Earth is now passing through a new carbon isotope maximum, it is of vital importance to understand the cyclic variations in the ocean carbon reservoir and its climate impact. The Pre-Quaternary variations in carbon and oxygen isotopes are characterized by their co-variations at the 400-ka eccentricity band, but the response of δ^{13}C and δ^{18}O to orbital forcing in the Quaternary became diverged with the growth of the Arctic ice-sheet. The present paper is the second summary report of ODP Leg 184 to the South China Sea.**

Keywords: ODP Leg 184, South China Sea, tropical forcing, carbon cycling, orbital periodicity.

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In the paper "Thirty Million Year Deep-Sea Records in the South China Sea"[1], we have reported the stratigraphic results of ODP Leg 184 to the South China Sea (SCS) in spring 1999, introduced the establishment of the best deep-sea stratigraphic sequences in the Western Pacific, and discussed changes in climate cyclicity over the past 20 Ma. As a continuation to the previous summary report, the present paper will focus on the long-term cycles in carbon cycling and their role in modulating the climate periodicity, the most significant results from our post-cruise studies of Leg 184.

One of the most outstanding contributions to paleoclimatology in the 20th century is the theory of orbital forcing or Milankovitch theory. The calculated changes in the Arctic solar radiation budget in summer at 65°N were found to be in agreement with glacial cycles in the geological records and, hence, responsible for the cyclicity in Quaternary glaciation. The scientists further noticed that the initial response has been propagated by variations in the North Atlantic Deep Water (NADW) production to the global ocean. Therefore, the Northern Hemisphere high latitudes have been taken as the source of the climate changes in Quaternary glacial cycles.

Precisely speaking, the Milankovitch theory has correctly pointed out the driving force of the Quaternary climate cyclicity, but does not yet identify the mechanism how the slow periodical changes in the Earth orbits lead to glacial cycles. A number of unsolved problems have arisen with the establishment of the Milankovitch theory which attributes the glacial cycles to variations in solar radiation received by the Northern Hemisphere high latitudes. For example, why the 100-ka periodicity predominates in the last 0.8—0.9 Ma, given the small eccentricity forcing in the 100-ka band ("100-ka problem")? Eccentricity has periodicities of 100-ka and 400-ka, but why a response at periods of 400-ka is missing from geological records ("400-ka problem")? How to explain the mismatch between the largest reduction in ice volume and a minor amplitude in the insolation change from MIS 12 to 11 about 400 ka ago ("Stage 11 problem")[2]? Besides, the insolation variations in the Northern Hemisphere alone cannot account for the approximately synchronous response to precession forcing which is out of phase between the two hemispheres[3]. Further problems have been emerging with new discoveries in recent years. Thus, the atmospheric concentration of CO_2 over the past 400-ka, as recorded in the Antarctic ice cores, varied in the same glacial cycles as ice-volume (δ^{18}O)[4], so how did the orbital cycles drive CO_2 variations? Comparisons of high-resolution records have revealed a time lag of Arctic behind Antarctic with respect to temperature changes in glacial cycles, a time lag of ice volume change behind that of CO_2[5], and a similar lag of high-latitude temperature changes behind those at middle and low latitudes. All these challenge the prevalent wisdom of Arctic control of the global climate system. In fact, something is missing in our understanding of the link between the Milankovitch astronomical forcing and the recorded climate changes[3], and the overlooked factor, as we believe, is the tropical forcing and carbon cycling.

The Global Changes studies in the last nearly twenty years have demonstrated the crucial role played by the tropical ocean in the global climate system. In the modern ocean, the Western Pacific Warm Pool (WPWP) receives a maximal amount of energy from solar radiation and exerts influence on the global climate through monsoon, ENSO and other low-latitude processes[6]. These air-sea exchanges can generate climate changes at high latitudes, an example is the tropical origin of the North Atlantic Oscillations (NAO)[7]. There is no reason for these features in the modern Earth's climate system to be absent in the geological past. The applications of new techniques have already invalidated the 25-year-old CLIMAP conclusion about stable SST in the tropics during glacial cycles[8], and the scientific focus now is being laid on the role of tropical Pacific in climate changes on orbital and millennial time scales[9]. Moreover, variations in CO_2 concentration are most probably the cause of the glacial cycles, rather than the consequences of ice-sheet changes[5].

In sum, the traditional Milankovitch theory is now facing challenges. Where is the key area in response to orbital variations to cause the glacial cycles: the Northern Hemisphere high latitudes, or the low latitude tropics? What drives climate cycles: the "ice-sheet forcing", "tropical forcing", or "CO_2 forcing"? The Arctic control concept has evoked heated debates. Obviously, there is a long way ahead before all the questions to be resolved, but the key is the geological record, and only this record can eventually test hypotheses. The records yielded by ODP drilling in the SCS, in particular the high-resolution long-term records from the tropical deep-water Site 1143 in the Nansha area, have provided valuable data for studying the variations in carbon cycling and tropical forcing. The present paper is an attempt to explore the issue of orbital forcing in climate cycles, on the basis of the new findings from Site 1143 and of comparisons with records from the global ocean.

1 Long-term periodicity of variations in ocean carbon reservoir

(ⅰ) Discovery of Carbon Isotope Maximum (δ^{13}Cmax). ODP Site 1143 is located in the southern SCS (9°21.72′N, 113°17.11′E, water depth 2772 m), within the Nansha or "Dangeroius Ground" coral reef area. The 510 m penetration at the site has reached the base of the upper Miocene about 12 Ma[10]. A total of 1992 samples taken from the upper section of 191 m were analyzed for isotopic composition of foraminifera, resulting in a 5-Ma continuous record with a resolution of 2—3 ka in average. After astronomical tuning, the Site 1143 sequence provides a first high-resolution, 5-Ma long record for the Western Pacific and one of the best such records from the global ocean[11]. For details of material, methods and calculations the readers are referred to our previous papers[11—13]. Here these long sequences of oxygen and carbon isotopes of planktonic *Globigerinoides ruber* and benthic *Cibicidoides* spp. are shown in Fig. 1. Noticeable

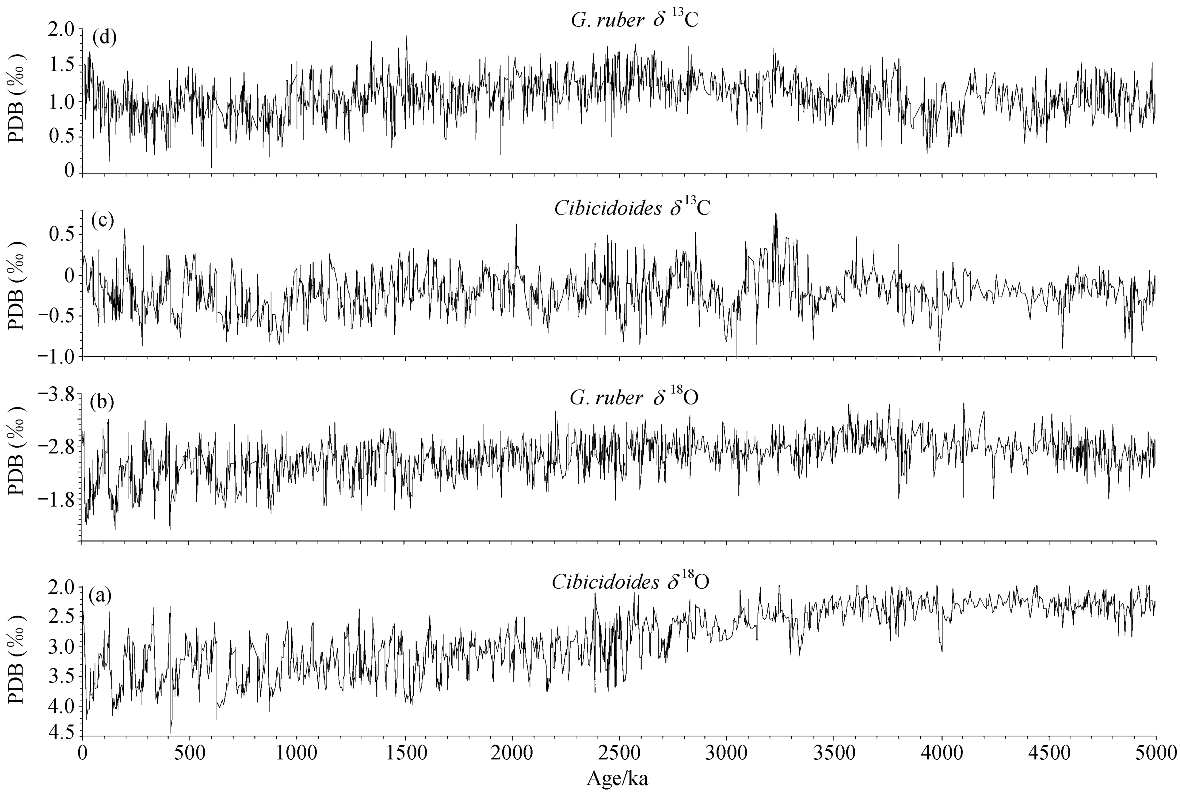

Fig. 1. 5 Ma isotope records from ODP Site 1143. (a) Benthic δ^{18}O of *Cibicidoides* spp; (b) planktonic δ^{18}O of *Globigerinoides ruber*; (c) benthic δ^{13}C of *Cibicidoides* spp.; (d) planktonic δ^{13}C of *Globigerinoides ruber*[11,13].

in the carbon isotope curves is the long-term cyclicity of 400—500 ka superimposed on high- frequent fluctuations. In other words, both the planktonic and benthic δ^{13}C sequences are punctuated by periodic occurrences of heavy values (carbon isotope maximum or "δ^{13}Cmax") with a baseline shift of ~0.3‰ (Fig. 1(c), (d)), although some cycles are more distinct than the others.

To ascertain whether the above described low-frequency cyclicity is local or global in nature, available long-term δ^{13}C sequences (>2 Ma in time length) from 8 sites over the global oceans are compared (Table 1). All the δ^{13}C profiles were correlated based on δ^{18}O stratigraphy before collectively compared as shown in Fig. 2. All the 9 sites across three oceans display the same low-frequency periodicity of 400—500 ka with well correlated δ^{13}Cmax events. For convenience in further discussion, the 9 events within the 4 Ma are labeled in a descending order as δ^{13}Cmax-Ⅰ to δ^{13}Cmax-Ⅸ (Table 2), with δ^{13}Cmax-Ⅰ being the one the Earth is currently experiencing.

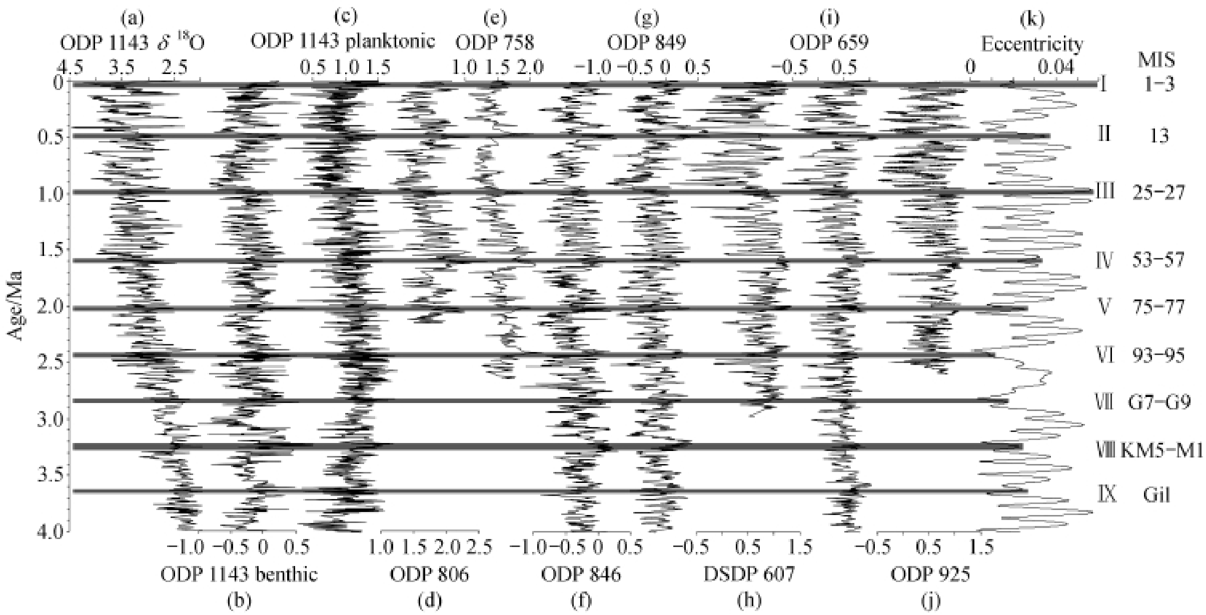

Fig. 2. A global correlation of δ^{13}C curves for the last 4 Ma (for locations and water depth see Table 1). Western Pacific: ODP 1143 (a) benthic δ^{18}O; (b) benthic δ^{13}C; (c) plankton δ^{13}C; (D) ODP 806 plankton δ^{13}C[14]. Indian Ocean: (e) ODP 758 plankton δ^{13}C[18]. Eastern Pacific: (f) ODP 846 benthic δ^{13}C [17]; (g) ODP 849 benthic δ^{13}C[16]. Northern Atlantic: (h) DSDP 607 benthic δ^{13}C[20]; (i) ODP 659 benthic δ^{13}C [21]]; (j) ODP 925 benthic δ^{13}C[22]; (k) eccentricity. Gray bars denote the carbon isotope maximum events (δ^{13}Cmax) showing 400—500 ka periodicity, and δ^{13}Cmax corresponds to eccentricity minimum (k) before the Quaternary. Ⅰ—Ⅸ on the right show the succession of δ^{13}Cmax.

Table 1 Long sequences of stable isotope records in the global ocean

Ocean	ODP site	Location	Water depth /m	Time interval /Ma	Sediment rate/cm • ka^{-1}	Resolution /ka	Foraminifera	Reference
West Pacific	806	0°19′N 159°22′E	2534	2.1	2.1	4—5	P	[14]
	1143	9°22′N 113°17′E	2772	5	3.9	2.6 (P) 2.8 (B)	P, B	this paper
East Pacific	677	1°12′N 83°44′W	3461	2.6	4.3	2.3	P, B	[15]
	849	0°11′N 110°31′W	3851	5	2.8	~4	B	[16]
	846	3°06′S 90°49′W	3296	6	4.2	2.5	B	[17]
Indian	758	5°23′N 90°21′E	2925	3.6	1.5	~7	P, B	[18,19]
North Atlantic	607	41°00′N 32°58′W	3427	2.8	4.5	~4	B	[20]
	659	18°05′N 21°02′W	3070	5	2.9	~4	B	[21]
	925	4°12′N 43°29′W	3041	2.6	3.2	3.1	B	[22]

P, Planktonic; B, benthic.

Table 2 Carbon isotope maximum events in the global ocean over the last 3 Ma

δ^{13}Cmax	Marine isotope stage (MIS)	Age/Ma
Ⅰ	1—3	0—0.05
Ⅱ	1—3	0.47—0.53
Ⅲ	25—27	0.95—1.00
Ⅳ	53—57	1.55—1.65
Ⅴ	75—77	2.00—2.06
Ⅵ	93—95	2.38—2.44
Ⅶ	G7—G9	2.73—2.80
Ⅷ	KM5—M1	3.20—3.29
Ⅸ	Gi1	3.56—3.61

(ⅱ) δ^{13}Cmax and glacial cycles. Comparison of the δ^{18}O and δ^{13}C over the last 1 Ma at Site 1143 (Fig.3) revealed a link between the δ^{13}Cmax events and the glacial cycles: Each δ^{13}Cmax was closely followed by major changes in glacial cyclicity and expansion of ice-sheet size[13]. Thus, δ^{13}Cmax-Ⅱ at MIS 13 ca. 500 ka ago was followed by a carbon shift that in turn led to a great expansion of ice-sheet at MIS 12/11 and the "Mid-Brunhes Event"[23]. Similarly, δ^{13}Cmax-Ⅲ about one million years ago and the subsequent carbon shift gave rise to the major glaciation of MIS 22 and the "Mid-Pleistocene Revolution"[24] when the 40-ka cyclicity was replaced by the 100-ka dominance. The details of these changes can be demonstrated with δ^{13}Cmax-Ⅱ as an example.

As seen from Fig. 3(c), the plankton δ^{13}C reached almost 1.5‰ at MIS 13 about 530—470 ka, then decreased to ~0.4‰ at the end of MIS 12 with a baseline shift about 0.3‰ and a significant increase in the amplitude of fluctuations, and returned back to the next carbon isotope maximum, δ^{13}Cmax-Ⅰ, at MIS 1-3. A similar succession occurred with the benthic δ^{13}C record, and the only difference is a drastic negative excursion of 0.8‰ during MIS 12 followed by a return to heavy values at MIS 11 (Fig. 3(b)), presumably associated with changes in the deep-water. Remarkable is the abrupt change in oxygen isotope succeeding the δ^{13}Cmax and carbon shift: The 2‰ negative shift in benthic δ^{18}O at the MIS 12/11 transition (Fig. 3(a)) is the largest-amplitude change in δ^{18}O of the global ocean over the past 6 Ma[25], indicating an extremely large-scale ice-sheet melting and sea level rising event. The sea level at the MIS 12/11 turn, about 430-ka, was by 20 m below the last glacial maximum lowstand, then suddenly rose up to 20 m above the present high-stand. This unusual change is ascribed to partial ice-sheet collapse in the Antarctic, leading to the "Mid-Brunhes Event". Because of the orbitally induced variations in insolation was minor, this radical event can not be accounted for by physical factors along ("MIS 11 problem", see above), and the solution of the problem most probably lies in the δ^{13}Cmax event at MIS 13 and the carbon shift at MIS 12[13].

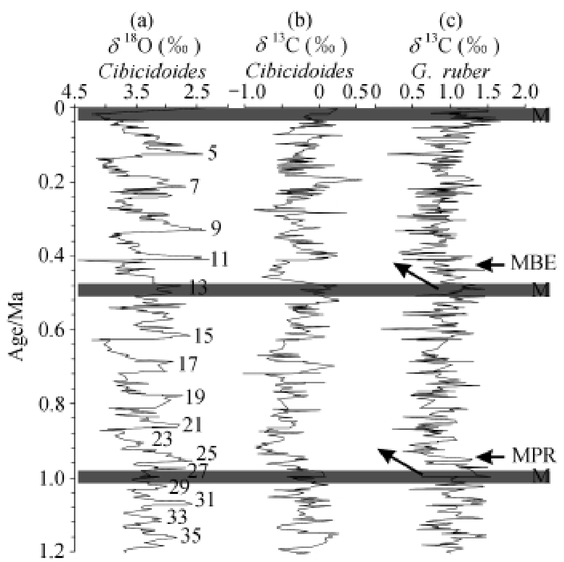

Fig. 3. Oxygen and carbon isotopic records spanning 1.2 Ma from ODP Site 1143. (a) Benthic δ^{18}O; (b) benthic δ^{13}C; (c) planktonic δ^{13}C. Numbers labeling δ^{18}O curve indicate marine oxygen isotope stages (MIS), grey bars denote carbon isotope maximum (δ^{13}Cmax), MBE stands for Mid-Brunhes Event, MPR for Mid-Pleistocene Revolution. Arrows show that δ^{13}Cmax preceded the transition in glacial cyclicity and expansion of ice-sheet[13].

What actually happened at the δ^{13}Cmax episode in the global ocean? Again we take MIS 13 as an example. In the equatorial Indian Ocean, an exceptional δ^{18}O depletion peak ("Event Y") at 525 ka in core MD 900963, equatorial Indian Ocean (5°03′N, 73°53′E, 2446 m w.d.) was followed by a heavy peak at 510 ka, "Event X", with a total amplitude of 2‰ (Fig. 4(f)). Event Y was interpreted as a large freshwater discharge to the surface ocean caused by abnormal monsoon precipitation and floods[26]. At the same time, a thick layer of sapropel dated 528—525 ka occurred in eastern Mediterranean (Fig. 4(e)), which has been ascribed to the unusually strengthened African monsoon[27]. All these findings are indicative of a large, regional-scale monsoon anomaly. Off the Amazon delta, in the equatorial Atlantic, abnormally enhanced terrigenous sediment flux at MIS 13 was recorded in ODP holes (Fig. 4(g)), implying unusually heavy precipitation and erosion in the Amazon drainage basin[28]. On the Asian continent, deepest weathering at MIS 13 was reported from the Chinese Loess Plateau, shown by the highest values of magnetic susceptibility of paleosol S5-1(Fig. 4(h)) due to the most intensive summer monsoon[29]. In sum, the δ^{13}Cmax-Ⅱ stage was distinguished by extremely warm and humid conditions in the entire tropical ocean and low latitudes, with intensified monsoon. The δ^{13}Cmax-Ⅱ event was followed by carbon shift and then an extra-large scale glaciation at MIS 12 about 430 ka (Fig. 3), implying that it acted as a major disturbance in the oceanic carbon reservoir and caused the major ice-sheet expansion and the "Mid-Brunhes Event" centred at MIS 11.

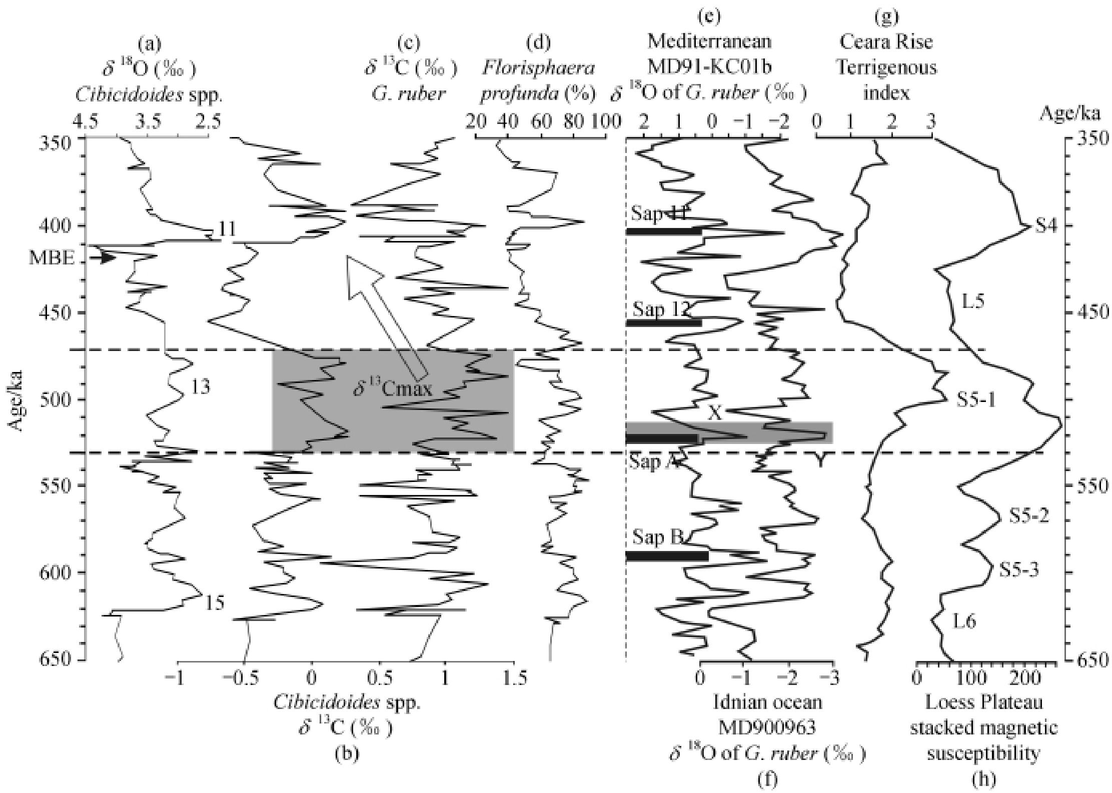

Fig. 4. Records of events around the δ^{13}Cmax-Ⅱ in MIS 13. ODP Site 1143, SCS: (a) Benthic δ^{18}O (MBE—Mid-Brunhes Event); (b) benthic δ^{13}C; (c) plankton δ^{13}C; (c) nannoplankton *Florisphaera profunda* %. Other oceans: (e) planktonic δ^{18}O record with sapropel layers (Sap) in core KC01b, Mediterranean (36º15′ N, 17º44′ E, w.d. 3643 m): "Sap A" is abnormal[27]; (f) planktonic δ^{18}O record in core MD 900963, equatorial Indian Ocean (5º03′ N, 73º52′ E, w.d. 2446 m), showing δ^{18}O events Y and X[26]; (g) "Terrigenous index" (relative terrigenous mass accumulation rate) at ODP Sites 929/925, equatorial Atlantic[28]; (h) stacked magnetic susceptibility from Loess Plateau, China (S, paleosol; L, loess)[29].

The records of carbon isotope maximum leading to ice-sheet expansion are not only restricted to δ^{13}Cmax-Ⅱ; a similar sequence of events has been found between δ^{13}Cmax-Ⅲ and the "Mid-Pleistocene Revolution"[13]. As shown in Fig. 3, the δ^{13}Cmax stage corresponding to MIS 25—28, about 0.95—1.00 Ma, was also followed by a negative carbon shift, leading to ice-sheet expantion. This was reconfirmed by a high-resolution study at ODP site 925, off the Amazon Delta[30]. Differing from δ^{13}Cmax-Ⅱ, the consequences of this event were not only the ice-sheet expansion, but also an extension of the glacial stage or a change in glacial cyclicity leading to the transition from 40-ka to 100-ka cyclicity, so called "Mid-Pleistocene Revolution"[24]. As to δ^{13}max-Ⅳ at MIS 53—57 about 1.65—1.55 Ma, the data available are mush more sparse. Nevertheless, it was found recently that the 100-ka cyclicity in climate might have started at 1.5—1.4 Ma[31], and that sea-ice expansion and a fundamental changes in South Ocean deep water circulation took place after MIS 52, at about 1.55 Ma[32] , all suggesting the impact of δ^{13}max-IV on the pacing of glacial cycles.

(iii) Reorganization of ocean carbon reservoir and major Quaternary stages in carbon perspective. The long-term eccentricity cycles, described above, had been recognized in earlier carbonate sediment studies. Twenty years ago, some 400—500 ka cycles coherent with the orbital eccentricity were found in the Quaternary and late Miocene carbonate records from the equatorial Pacific[33]. Later, long-term aragonite cycles with approximately 500 ka wavelength were established in the Atlantic and Indian Ocean (Fig. 5(d))[34]; high-amplitude oscillations with a periodicity of ~500 ka in coarse fraction records were reported from the tropical Indian Ocean (Fig. 5(c)) [35], and similar oscillations also occur at Site 1143, SCS (Fig. 5(b))[36]. These discoveries all correspond to the low-frequency cycles in carbon isotope (Fig. 5(a)). At least in the Indo-Pacific, the 500 ka periodicity has been observed in a wide range of water depths, from 540 m (ODP Site 716, Indian Ocean) to near 4500 m (V24-58, Equatorial Pacific). Therefore, it must be attributed to changes in oceanic chemistry throughout the water column, rather then to deep-water carbonate dissolution caused by changes in deep-water circulation.

Both carbon isotope and carbonate sediment are archives of the history of carbon cycling and carbon reser-

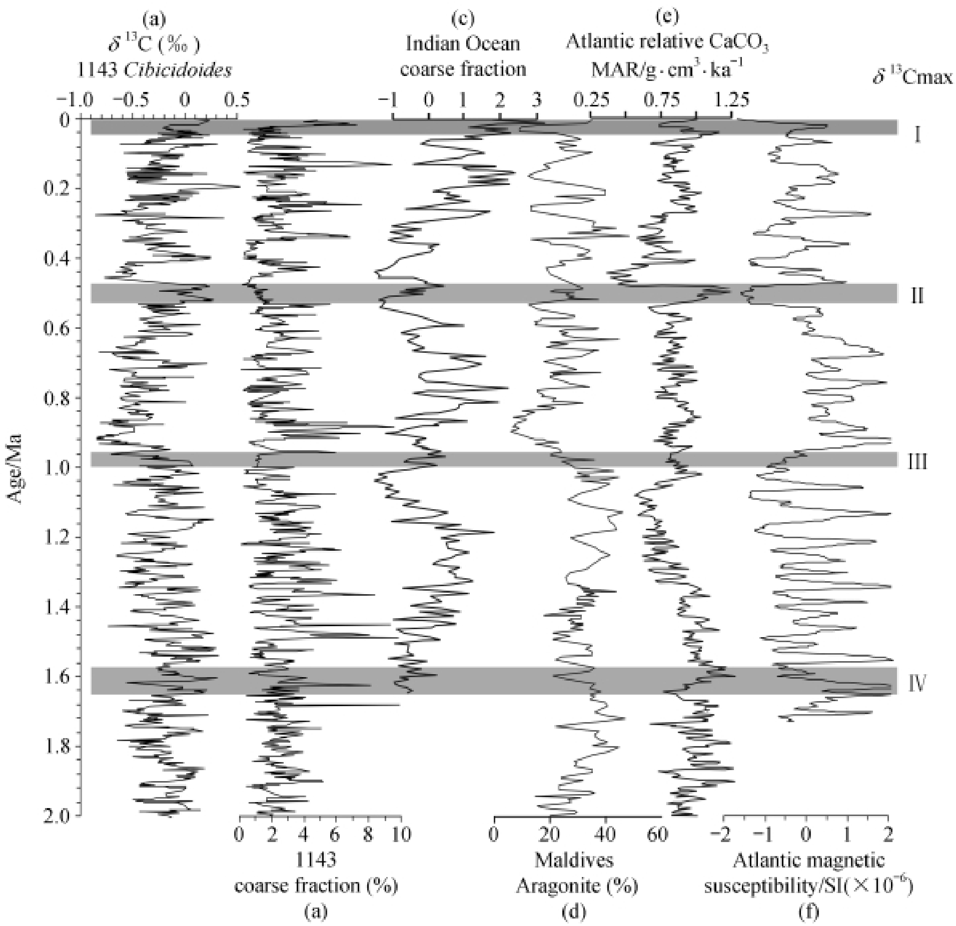

Fig. 5. Correlation of oceanic δ^{13}Cmax with carbonate preservation variations over the past 1.3 myr. (a) ODP 1143 benthic δ^{13}C; (b) ODP 1143 coarse fraction% (>63 μm); (c) composite coarse fraction index, tropical Indian Ocean[35]; (d) fine aragonite % at ODP Site 716B (4º56′ N, 73º17′ E, w.d. 540 m)[34]; (e) relative $CaCO_3$ mass accumulation rates at ODP Sites 926/925[28]; (f) Subtropical South Atlantic magnetic susceptibility stack (SUSAS)[36].

voir in the ocean. The carbon reservoir in the ocean water is nearly 60 times of that in the atmosphere, and dissolved inorganic carbon in the ocean water is controlled by its alkalinity and presented as HCO_3^-, CO_3^{2-}, CO_2 in varying proportions. Sea water exchanges its carbon with the atmosphere and the lithosphere through a variety of processes including “biological pump”, “carbonate pump”, degassing, deep-water dissolution of carbonate and others. These processes are crucial in the global carbon cycling, and can have their imprints preserved in the sea water δ^{13}C and in carbonate sediments. On geological time scales, the variations of oceanic δ^{13}C are mainly determined by the ratio between organic vs inorganic component in the carbon flux. As inorganic carbon is basically represented by carbonate, the coeval variations of δ^{13}C and carbonate in the 500-ka periodicity denote reorganization in the oceanic carbon reservoir which must have its impact on carbon cycling in the entire Earth system process. It can not be merely by coincidence that the δ^{13}Cmax-Ⅱ event 500 ka ago has brought about not only the major ice volume increase, but also the “Mid-Brunhes” event centred at MIS 11 about 400-ka, an event of deep-sea carbonate dissolution[23].

A common practice in Quaternary climate history studies is just to pursue ice-volume variations as exhibited by δ^{18}O, by considering carbon system changes shown by δ^{13}C as the consequences of ice-volume changes. If some periodicity was found in the δ^{13}C record, it remained difficult to be linked to orbital forcing[22]. The new finding at Site 1143 in the SCS has offered a new aspect to understanding the Quaternary history: variations in carbon system as revealed by δ^{13}C are not only passively responding to the ice-sheet changes; rather, carbon cycling and carbon reservoir themselves are subject to orbital forcing, resulting in their own periodicity and variations. From a carbon perspective, therefore, the Quaternary period has passed through three major stages defined by four δ^{13}Cmax events, and each appears to represent a further step in ice-cap development (Fig. 6; Wang et al., MS[1)]). Interest-

1) Wang, P., Tian, J., Chen, X. et al., Major Pleistocene Stages in a Carbon Perspective: The South China Sea record and its global comparison (MS).

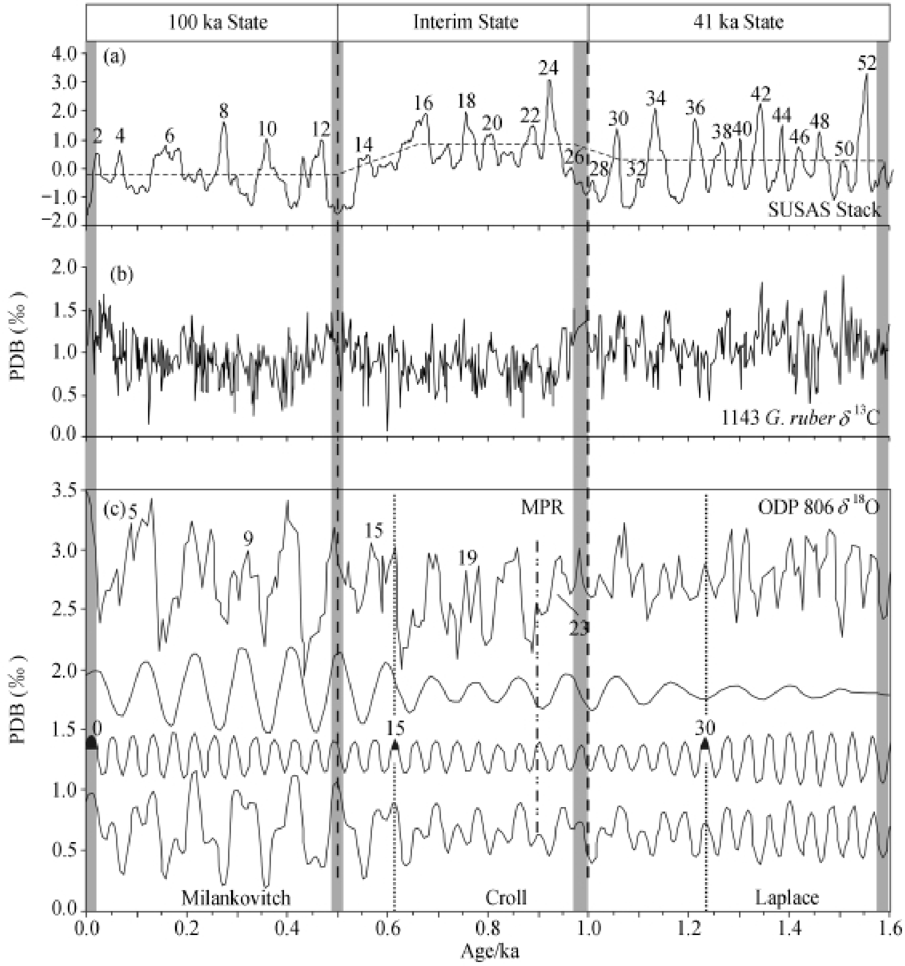

Fig. 6. Subdivision of the last 1.6 Ma of the Quaternary into major climate stages based on the evolution of oceanic carbon system. (a) Triple division into "100 ka State","Interim State" and "41 ka State" based on Subtropical South Atlantic Susceptibility (SUSAS) stack[36]; (b) triple division based on four δ^{13}Cmax events, represented by planktonic δ^{13}C of ODP Site 1143, SCS; (c) triple division into "Milankovitch", "Croll", and "Laplace" chrons based on predominant climate cyclicity represented by benthic δ^{18}O of ODP 806, western tropical Pacific[24].

ingly, the triple division discussed here coincides with the outcome of the South Atlantic studies. From magnetic susceptibility logs of subtropical South Atlantic deep-water cores, Schmieder et al. (2000) argued that the "Mid-Pleistocene Revolution" should be regarded as a specific, transitional stage, and in terms of climate transition the Quaternary comprises three states: the early "41 ka state", the "interim state" and the late "100 ka state"(Fig. 6(a)), with the same age boundaries as in our δ^{13}C record (Fig. 6(b)). There was still an earlier triple division proposed by W. Berger et al. (1993) who divided the 1.8-Ma history of the Quaternary into "Milankovitch", "Croll" and "Laplace" chrons, each with 15 obliquity cycles or 600 ka in duration (Fig. 6(c)). This latter scheme also depicts the progressive change in glacial cyclicity, but the age boundaries seem to be rather arbitrary.

(ⅳ) Long-term periodicity in Pre-Quaternary ocean carbon system. As seen from a comparison of the δ^{13}C curves at various sites (Fig. 2 (b)—(j)) with the eccentricity (Fig. 2(k)), each of the pre-Quaternary carbon isotope maxima occurred at a long-term eccentricity minimum on about 400 a spacing before the last million years, but the δ^{13}Cmax events in the Quaternary were separated by a 500-ka interval without matching with the eccentricity minimum. To understand the change, a much longer record of the δ^{13}C history is needed.

The oldest record of high-resolution isotope sequence based on astronomically tuned time scale is available from an interval across the Oligocene/Miocene boundary. The benthic isotope records over a 5-Ma interval (20.5—25.4 Ma) from ODP Site 929, western equatorial Atlantic, exhibit low frequency cycles of 400 ka in both δ^{18}O and δ^{13}C records (Fig. 7(a)). Most of the δ^{13}Cmax events are correlated with positive δ^{18}O excursions, showing coherence between δ^{18}O and δ^{13}C records, as well as with the eccentricity minimum. This indicates a

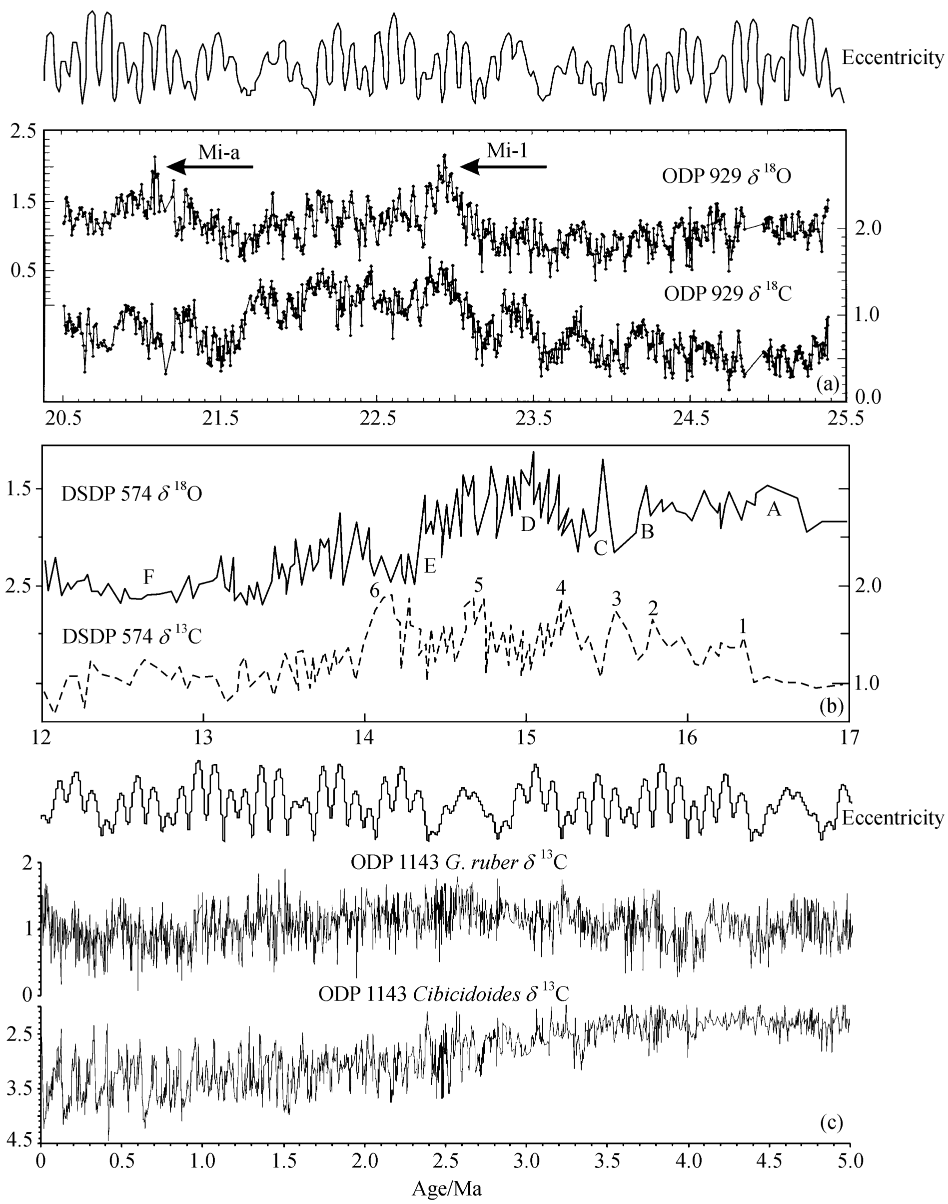

Fig. 7. Comparison of deep-sea benthic δ^{18}O and δ^{13}C records over three time intervals in the late Cenozoic: (a) Late Oligocene-early Miocene, 20.5—25.4 Ma, ODP Site 929, western equatorial Atlantic (5°58′N, 43°44′W, w.d. 4358 m), compared with orbital eccentricity[37]; (b) Middle Miocene, 12—17 Ma, DSDP Site 574, eastern equatorial Pacific (4°12′N, 133°19′W, w.d. 4561 m), A—F denote δ^{18}O events, 1—6 denote δ^{13}Cmax events CM1-CM6[39]; (c) Plio-Pleistocene, 0—5 Ma, ODP Site 1143, southern SCS, compared with orbital eccentricity.

simple and consistent relationship between carbon cycling and glacial cycles due to effective eccentricity forcing[37,38]. Such a relationship maintained in the middle Miocene. From 17 Ma to 13.5 Ma, six δ^{13}Cmax (CM1-CM6) occurred at intervals of about 440 ka suggestive of the 413 ka eccentricity cycles, correlating again with positive δ^{18}O excursions (Fig. 7(b))[39].

The covariance and correlation between oxygen and carbon isotopes at the long-term eccentricity band persisted until the Pliocene and remained evident at least 3 Ma ago (Fig. 7(c)). This is believed to be the characteristic of Antarctic glacial cycles. With the development of Arctic ice sheet, the relation between oxygen and carbon isotopes in long-term cycles has been complicated: Over the last million years, the 100-ka periodicity turned predominant and the 400-ka vanishes in the δ^{18}O record but extended to ~ 500-ka in the δ^{13}C record. This explains the long neglect of the 400-ka cycles by the Milankovitch theory focusing on the late Quaternary. The continuous isotope sequence of 24 Ma from ODP Site 1148, northern SCS (18°50′N, 116°34′E, w.d. 3294 m)[1], covers all three sections shown in Fig.1. A comparison between its δ^{18}O and

δ^{13}C records provides the same features as discussed above; and the 400-ka long-term periodicity throughout the δ^{13}C sequence has remained stable and become most prominent during the last 3 Ma, in contrast to the strongly fluctuating δ^{18}O record.

To sum up, the δ^{18}O and δ^{13}C covaried at the eccentricity bands before the Quaternary, but their response to orbital forcing diverged in the last million years with the growth of the Arctic ice cap. Therefore, the last 1 Ma was "abnormal" for the Cenozoic, and the close correspondence between δ^{13}C cycles and eccentricity was "normal" in longer geological records. In other words, the late Cenozoic has witnessed two types of δ^{13}Cmax: Before the Quaternary, δ^{13}Cmax occurred at a 400-ka interval, and both δ^{13}C and δ^{18}O displayed their positive excursion at the eccentricity minimum in 400-ka cycles, implying the response of carbon cycling to orbital forcing consistent with glacial cycles at the eccentricity band. During the Quaternary, the periodicity of δ^{13}Cmax extended to 500-ka, and its occurrence ceased to correspond with eccentricity minimum or positive excursion in δ^{18}O; instead, each δ^{13}Cmax event led to transition in glacial cyclicity and major increase of ice volume[13]. In the following, we will argue the origin of periodical δ^{13}Cmax occurrences from tropical processes as a low-latitude response to orbital forcing. The above-discussed changes in the Quaternary probably denote different relationships between the Arctic *vs* Antarctic ice-sheets and tropical processes: The Antarctic has an open connection and hence a simpler link with the tropical ocean, whereas the continents that surround the Arctic hamper its connection with the tropical ocean, causing a much more complicated relationship between them.

2 Tropical forcing in climate evolution

(ⅰ) Tropical forcing and its periodicity. As shown above, the carbon isotope changes do not passively follow those of oxygen isotope, and the oceanic carbon reservoir has its own periodicity of variations. Then, which process is responsible for the long-term cyclicity of 400—500 ka in carbon cycling? The heavily increased precipitation over low latitudes at δ^{13}Cmax-Ⅱ or MIS 13 (see 2(ⅱ), Fig. 4) suggests a possible connection of δ^{13}Cmax with tropical processes. Characteristic features of periodicity in geological records may provide some useful indication for discriminating high- from low-latitude forcing.

There are geographic differences in responding to orbital parameters. The precession forcing (20-ka cycles) dominates low latitudes, but the effects of obliquity (40-ka cycles) are more evident at higher latitudes. These two cyclicities roughly correspond to "monsoon response" and "ice sheet response" to orbital forcing, respectively[40]. Because the amplitude of variability in climate effects of precession (sin ω) is modulated by eccentricity (ε) ("climate precession", ε sin ω), the 100- and 400-ka eccentricity together with the 20-ka precession characterize the low-latitude processes[41]. In addition, there are also "semi-precessional cycles" of about 10 ka in the tropics, as the sun passes there directly overhead twice in a year[42,43]. As a result, low-latitudes have the richest spectrum of climate response to orbital forcing[44], ranging from 10-ka to 400-ka and even 2000-ka. This is exactly what we found from spectral analyses of planktonic δ^{13}C record at Site 1143. The 400-ka periodicity of δ^{13}Cmax, the rich spectrum of δ^{13}C record from the tropical surface water and the abnormal climate changes at δ^{13}Cmax in the tropics during the Quaternary, all together show a close tie between tropical processes and reorganization of ocean carbon reservoir.

As the tropics must have been the main source of climate forcing in an ice-free "Hot-House" world, tropical forcing of climate changes becomes even more prominent when a longer geological history is considered. Indeed, the 400- and 100-ka eccentricity cycles have been reported from late Triassic rhythmic deposits of tropical lakes (Newark Group)[45], from the carbonate content in middle Cretaceous Albian shelf deposits in Germany[46], and from the Ca/Fe ratio in late Eocene deposits in North Atlantic[47]. The 20-ka precessional cycles have also been reported from widely distributed locations. Therefore, the role of obliquity was subordinate in a "Hot-House" world[48], but increased with the growth of the Arctic ice cap, as a characteristic of the high-latitude response. Meanwhile, precession and eccentricity have maintained their dominating role in forcing tropical processes like monsoon, and the 400-ka cycles have been evident in monsoon-induced sapropel layers in the Mediterranean[49], in eolian dusts in the equatorial Atlantic[21], indicating a modulation of glacial cycles through by these long-term cycles in the carbon system.

Among various orbital parameters, the 400-ka eccentricity is considered as most significant. Because of the presence of tropical forcing throughout the Hot-House and Ice-House regimes and because of the astronomical stability of the long-term eccentricity[47], the 400-ka periodicity may become the "pendulum" or "tuning fork" in geological timing. This applies especially to the early geological past when the recognition of precession signal itself may be hampered by a limited time resolution, the 400-ka eccentricity suits best for geological time calibration[50].

(ⅱ) Diatom and carbon reservoir. A working hypothesis. We have demonstrated the connection of δ^{13}C with tropical processes. However, it remains enigmatic what actually happens with the ocean carbon reservoir through a δ^{13}Cmax episode, why its occurrences corresponded to the 400-ka eccentricity cycles, and how its periodicity extended to 500-ka in the last million years. As a working hypothesis on available data, we suggest that long-term variations in chemical weathering that changed the silicon supply from the continent to the ocean have determined the diatom/coccolith ratio in phytoplankton and subsequently the organic/inorganic ratio in the sinking particles sinking onto the ocean floor.

It has been well established that the effect of oceanic "biological pump" in controlling CO_2 in atmosphere depends on the composition of phytoplankton. The two major microfossil groups in geological records, diatoms and coccoliths, play very different roles in carbon cycling: diatoms produce only organic carbon, but coccoliths create also carbonate skeleton. They consume atmospheric CO_2 while producing organic carbon, but release CO_2 when producing carbonate. Therefore, variations in the ratio of the two groups in phytoplankton will effect the carbon cycling. On the other hand, oceanic $\delta^{13}C$ is related to the ratio between organic *vs* inorganic (carbonate) carbon burial rate, and $\delta^{13}C$ values will be changed by the diatom/coccolith ratio. This is the background for the recent advanced "Silica Hypothesis"[51], claiming that diatom production can alter atmospheric CO_2 level. In turn, the diatom/coccolith ratio depends on Si flux from the continent. When Si is available, diatoms flourish while coccoliths decline, subsequently raising the organic/inorganic carbon ratio in oceanic "sediment rain" and reducing the atmospheric CO_2 concentration[52]. According to estimation, a 40% decline in calcite (coccolith) flux from the ocean mixed layer would be sufficient to reduce atmospheric CO_2 from 280 ppm to 200 ppm[53]. During the glaciation, enhanced transport of eolian dust increased Si input to the ocean, leading to a decrease of coccolith flux[54]. However, the primary source of silicon in the modern ocean is from river input, which at present supplies Si from low latitude lands to the ocean over ten times more than the glacial eolian record[55]. Meanwhile, the raised diatom productivity may also lead to heavier values of oceanic $\delta^{13}C$. In result, the secular variations of low-latitude summer monsoon and continental weathering over 400-ka eccentricity cycles may have given rise to long-term changes in Si supply and diatom production, which in turn cause reorganization of ocean carbon reservoir, expressed as δ^{13}Cmax. If the hypothesis is tenable, the tropical response to orbital forcing can enter the global climate system through carbon cycling and modulate ice-volume at high latitudes[13].

In fact, the secular change in chemical weathering caused by eccentricity cycles has been considered as a possible mechanism influencing the long-term $\delta^{13}C$ trend in glacial cycles over the last 200 ka, when the amplitude of $\delta^{13}C$ variations decreased[56,57]. A recent geochemical study at ODP site 1145, SCS, has shown the existence of 400-ka cyclicity in chemical weathering[58]. At Site 1143, the subsurface planktonic foraminifera *Pulleniatina obliquiloculata* % displays its minimum at δ^{13}Cmax-Ⅰ, Ⅱ, and the subsurface nannoplankton *Florispheara profunda* % declines remarkably beginning from δ^{13}Cmax-Ⅱ (Fig. 4(d)), suggesting major changes in the ocean upper structure and in plankton[59]. All these observations support our hypothesis. As ocean water is unsaturated in Si, biogenic opal generated by diatoms is basically dissolved in sea water and can hardly be preserved in deposits. However, some "gigantic" diatom species can suddenly bloom in oligotrophic ocean water and accumulate as monospecific diatom layers. Occurrences of these diatom deposits have been reported from various oceans[36, 60, 61], underscoring the possible role of diatom in carbon cycling. Certainly, the working hypothesis at this stage is no more than a speculation, and much more further work is needed to develop and testify it.

(iii) Double forcing of glacial cycles. A major contribution from the ODP Leg 184 studies is the progress in our understanding of glacial cycles. In the previous summary report, we have pointed out the necessity of long sequences for investigating the climate cyclicity and demonstrated on the over 20 Ma long record at Site 1148 that the climate response to orbital forcing changed with the growth of ice cap, and the prevalent understanding of Milankovitch cycles in the late Quaternary is specific to an unusual scenario of the Earth system with both poles ice-caped[1]. On the basis of the high-resolution 5-Ma record at Site 1143, the present paper describes the existence of long-term periodicity in oceanic carbon cycling as a new insight into the link between carbon reservoir and ice cap, promoting a holistic approach to orbital forcing of the climate system. When the connection between $\delta^{18}O$ and $\delta^{13}C$ is considered in a perspective of late Cenozoic long sequence, the specifics of the Quaternary cyclicity in climate changes become apparent. The glacial cycles have been caused by a combination of response from high and low latitudes to orbital forcing, by way of physical processes (ice-sheet) and biogeochemical processes (carbon cycling), or, in short, by a "double forcing".

Therefore, it is a real bias to attribute the Quaternary glacial cycles solely to the insolation at the Northern Hemisphere high latitudes, and to restrict the role of low latitudes to a passive respondent. The "double forcing" concept, on the contrary, is helpful in resolving some "teethed problems" in the traditional version of the Milankovitch theory. Thus, the insolation change at the MIS 12/11 turn had been much too weak to account for the major change in ice-volume, but the discovered δ^{13}Cmax-Ⅱ event at MIS 13 which led to carbon shift and ice-cap expansion at MIS 12[13], provides an answer to the "Stage 11 problem". In contrast to the prominent 100-ka cycles, 400-ka cycles were obscure in glacial records; now the discovery of 400-ka periodicity in carbon cycling and its link with oxygen isotope indicates these long-term cycles have been driving the tropical climate and modulating the glacial cyclicity, a promising solution for the "400-ka problem". Besides, the waning and waxing of the 100- and 400-ka eccentricity periodicity over the 20 Ma have offer a new vision to address the "100-ka problem". The concept of tropical forcing explains the approximate synchronism of precession cycles acting on the two Hemispheres. Our evidence for modulation of glacial cycles by low-latitude forcing and carbon system is well in line with the recent discoveries of a lead in CO_2 and temperature changes at low and middle latitudes over ice-sheet.

In addition, the "double forcing" concept discussed above is significant because it deals with some fundamental issues in the prediction of future climate environment for the human society. Using the same Milankovitch theory, scientists warned 30 years ago that "the present-day warm epoch will terminate relatively soon"[62], but now it is conjectured that the current interglacial "may last another 50000 years"[63]. The opinions remain controversial and puzzling the public. The divergence in views roots in different estimations of the natural trend in carbon cycling change: Some scientists assume the declining late Cenozoic CO_2 level will continue and believe a new glaciation is coming[64]; but others consider CO_2 variations in the next 130-ka may only repeat what in the past 130-ka and conclude a new glaciation will start 50 ka from now (Fig. 8)[63]. None of the two has taken into account the periodicity in carbon cycling itself, and the presumed CO_2 trends are groundless in both cases. Noticeably, the Earth system is now entering a new eccentricity minimum, with the precession amplitude diminishing[63]. The oceanic carbon system is again passing through a new δ^{13}Cmax event (Fig. 3), and abnormal monsoon climate has already been observed in MIS 3[65]. It will be impossible to scientifically

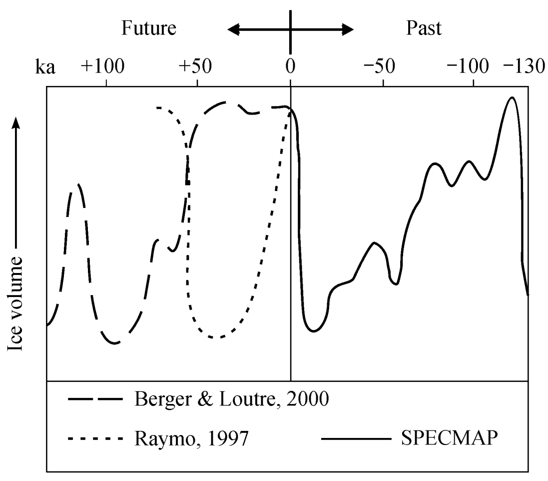

Fig. 8. Different versions of prediction for the future glaciation (human impact is excluded). The next glaciation is coming, according to Raymo, 1997 [64]; or the present interglacial will last other 50 ka, according to Berger & Loutre, 2002[63].

estimate the future long-term trend of the CO_2 level and climate, if these great changes in the carbon system remain ignored. This is to say, scientific prediction on future climate change will be unrealistic without a better understanding of the natural tendency of CO_2 level changes and the link between carbon system and glacial cycles.

3 Concluding remarks

The SCS ODP studies have significantly improved our understanding of the oceanic carbon system, including the periodicity of its variations and its role in glacial cycles. We found:

(1) The oceanic δ^{13}C records display a 400-ka long-term periodicity expressed as repeated δ^{13}Cmax episodes. δ^{13}Cmax events before the Quaternary corresponded to the eccentricity minima, but changed in the Quaternary when the 400-ka cycles extended to 500-ka.

(2) The δ^{13}Cmax events were accompanied by reorganization of the ocean carbon system, though the mechanism remains unclear. Judging from the intensified precipitation widely recorded at low latitudes during δ^{13}Cmax-Ⅱ about 500-ka ago, the reorganization is interpreted to have originated from low-latitude processes, probably through enhanced chemical weathering which led to proportional increase of diatoms among oceanic phytoplankton.

(3) In the Quaternary, δ^{13}Cmax was followed by major expansion of ice sheet or even the transition in glacial cyclicity. This indicates the low-latitude processes can modulate ice-sheet variations through the carbon system, and changes in the carbon system have their own periodicity. As the Earth is now passing through a new δ^{13}Cmax, it is impossible to scientifically predict the long-term trend of climate change without understanding the changes in the carbon system.

(4) As shown by many late Cenozoic δ^{18}O and δ^{13}C long sequences, the carbon system and ice sheet co-varied at the 400-ka band, implying a coordination between the Antarctic ice cap and low-latitude processes. In the Quaternary, δ^{18}O and δ^{13}C sequences diverged with the growth of the Arctic ice sheet, suggesting a complicated relationship between the Arctic ice cap and tropical processes.

(5) The traditional version of the Milankovitch theory is based on specific conditions of the late Quaternary and overestimates the role of the Northern Hemisphere high latitudes. Glacial cycles are, actually, generated by "double forcing", a combination of the responses of high and low latitudes to the orbital forcing, through physical processes of the ice cap and biogeochemical processes of carbon cycling.

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