Integrated “plume winter” scenario for the double-phased extinction during the Paleozoic–Mesozoic transition: The G-LB and P-TB events from a Panthalassan perspective

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Abstract

The event across the Paleozoic–Mesozoic transition involved the greatest mass extinction in history together with other unique geologic phenomena of global context, such as the onset of Pangean rifting and the development of superanoxia. The detailed stratigraphic analyses on the Permo-Triassic sedimentary rocks documented a two-stepped nature both of the extinction and relevant global environmental changes at the Guadalupian–Lopingian (Middle and Upper Permian) boundary (G-LB, ca. 260 Ma) and at the Permo-Triassic boundary (P-TB, ca. 252 Ma), suggesting two independent triggers for the global catastrophe. Despite the entire loss of the Permian–Triassic ocean floors by successive subduction, some fragments of mid-oceanic rocks were accreted to and preserved along active continental margins. These provide particularly important dataset for deciphering the Permo-Triassic paleo-environments of the extensive superocean Panthalassa that occupied nearly two thirds of the Earth’s surface. The accreted deep-sea pelagic cherts recorded the double-phased remarkable faunal reorganization in radiolarians (major marine plankton in the Paleozoic) both across the G-LB and the P-TB, and the prolonged deep-sea anoxia (superanoxia) from the Late Permian to early Middle Triassic with a peak around the P-TB. In contrast, the accreted mid-oceanic paleo-atoll carbonates deposited on seamounts recorded clear double-phased changes of fusulin (representative Late Paleozoic shallow marine benthos) diversity and of negative shift of stable carbon isotope ratio at the G-LB and the P-TB, in addition to the Paleozoic minimum in $^{87}$Sr/$^{86}$Sr isotope ratio in the Capitanian (Late Guadalupian) and the paleomagnetic Illawarra Reversal in the late Guadalupian. These bio-, chemo-, and magneto-stratigraphical signatures are concordant with those reported from the coeval shallow marine shelf sequences around Pangea. The mid-oceanic, deep- and shallow-water Permian records indicate that significant changes have appeared twice in the second half of the Permian in a global extent. It is emphasized here that everything geologically unusual started in the Late Guadalupian; i.e., (1) the first mass extinction, (2) onset of the superanoxia, (3) sea-level drop down to the Phanerozoic minimum, (4) onset of volatile fluctuation in carbon isotope ratio, $^{87}$Sr/$^{86}$Sr ratio of the Paleozoic minimum, (6) extensive felsic alkaline volcanism, and (7) Illawarra Reversal.

The felsic alkaline volcanism and the concurrent formation of several large igneous provinces (LIPs) in the eastern Pangea suggest that the Permian biosphere was involved in severe volcanic hazards twice at the G-LB and the P-TB. This episodic magmatism was likely related to the activity of a mantle superplume that initially rifted Pangea. The supercontinent-dividing superplume branched into several secondary plumes in the mantle transition zone ($410$–$660$ km deep) beneath Pangea. These secondary plumes induced the decompressional melting of mantle peridotite and pre-existing Pangean crust to form several LIPs that likely caused a “plume winter” with global cooling by dust/aerosol screens in the stratosphere, gas poisoning, acid rain damage to surface vegetation etc. After the main eruption of plume-derived flood basalt, global warming (plume summer) took over cooling, delayed the recovery of biodiversity, and intensified the ocean stratification. It was repeated twice at the G-LB and P-TB.

A unique geomagnetic episode called the Illawarra Reversal around the Wordian–Capitanian boundary (ca. $265$ Ma) recorded the appearance of a large instability in the geomagnetic dipole in the Earth’s outer core. This rapid change was triggered likely by the episodic fall-down of a cold megalith (subducted oceanic slabs) from the upper mantle to the D’ layer above the 2900 km-deep core-mantle boundary, in tight association with the launching of a mantle superplume. The initial changes in the surface environment in

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1. Introduction

The greatest mass extinction in the Phanerozoic marks the major era boundary between the Paleozoic and Mesozoic across which the Earth’s biosphere drastically changed from the older Paleozoic regime to the Mesozoic–Modern one (e.g. Sepkoski, 1984; Erwin, 1993, 2006; Fig. 1A). A huge proportion of Paleozoic marine and terrestrial invertebrates was terminated and the Mesozoic pioneer taxa filled the empty ecospace in the aftermath. The Paleozoic–Mesozoic boundary or Permian–Triassic boundary (P-TB) is outstanding among the Big 5 extinction-related boundaries in the Phanerozoic, not only in the greater magnitude of mass extinction per se (over 80% diversity loss on generic level) but also in recording quite unique geologic phenomena that are rarely recognized in the rest; i.e., chert gap, reef gap, coal gap, sea-level minimum, zenith of Pangea, prolonged anoxia, and volatile changes in various isotope ratios. Nonetheless, the ultimate cause is still the center of discussion (e.g. Erwin, 2006; Bottjer et al., 2008).

The double-phased nature of the end-Paleozoic mass extinction has been much emphasized during the last decade (e.g. Isozaki et al., 2004; Racki and Wignall, 2005; Isozaki, 2007b; Retallack et al., 2006); however, before 1994 not much attention was paid to the Middle-Late Permian boundary (or Guadalupian–Lopingian boundary; G-LB) event by the great shadow of the widely-recognized P-TB event. This was partly due to the lesser awareness among geologists outside China where the G-LB event is best documented. Since the first claim made by Jin et al. (1994) and Stanley and Yang (1994), the largest massacre in the Phanerozoic has been understood as a combined effect of two independent mass extinction events that took place back to back within a short-time period; i.e., first at the G-LB (ca. 260 Ma; Gradstein et al., 2004) and second at the P-TB sensu stricto (ca. 251–252 Ma; Bowring et al., 1998; Mundil et al., 2004) (Fig. 1B). The interval of nearly 8 million years between the two independent extinction events was probably too short for various Late Permian biota to retrieve their original high diversity and population size (Stanley and Yang, 1994). This double-phased extinction pattern suggests that unusually harsh conditions appeared at least twice in the Permian shallow seas around Pangea.

The G-LB has profound implications that are quite distinct from those of the P-TB. First the G-LB marks the initial major decline after the Late Paleozoic long-term biodiversity stability in a high plateau since the end-Devonian extinction, as clearly shown in Fig. 1A. Second, the G-LB extinction apparently coincided in timing not only with the onset of the long-term oxygen-depletion in ocean (superanoxia; Isozaki, 1997; Fig. 1B), but also with the lowest sea level in the Phanerozoic (Hallam and Wignall, 1999; Miller et al., 2005; Haq and Schutter, 2008) and the initial breakup of Pangea (Isozaki, 2007b) (Fig. 2). Third, the isotopic systems of carbon and strontium had started to change drastically in the Late Guadalupian (Veizer et al., 1999; McArthur and Howarth, 2004; Isozaki et al., 2007a,b; Kani et al., 2008), and kept fluctuating during the Late Permian and Early Triassic (e.g. Holser et al., 1989; Baud et al., 1989; Payne et al., 2004; Horacek et al., 2009) (Fig. 2). These varied geologic phenomena suggest that the main story of the Paleozoic–Mesozoic transition was not simple, like the end-Cretaceous (K-T boundary) scenario, such as with one large bolide impact and the sudden great mass-death, but was a long-term complicated event composed of at least two sub-stories and spanned for over 10 million years. A major reorganization of the global oceanographic regime likely started in the late Middle Permian, and the perturbation in biosphere culminated at the P-TB to drive the biggest biodiversity loss in the Phanerozoic (Fig. 1B).

In addition to the conventional studies on continental shelf sediments in S. China and peri-Tethyan regions, the double-stepped faunal turnover of fusulinids and radiolarians was recently documented at the G-LB and the P-TB also in the mid-superocean (Ota and Isozaki, 2006; Isozaki, 2007a), and these results suggest the global context of the two distinct events. This article at first reviews the current knowledge on the mid-oceanic records spanning across both G-LB and P-TB derived mostly from Japan, with respect to the double-phased extinction and relevant environmental changes of global context. Then, on the basis of all data for the double-phased changes in mid-Panthalassa and peri-Pangean domains, their geological implications are discussed with regard to the Permian mantle superplume activity. At the end, an integrated version of the “plume winter” scenario is introduced for explaining the cause, processes, and effects of the greatest biosphere catastrophe of the Phanerozoic that took place in the Paleozoic–Mesozoic transition interval (ca. 265–240 Ma).

2. Mid-oceanic records

The oldest extant ocean floor has an age of about 200 Ma (Early Jurassic), which is exposed in the deep-sea floor just off the Mariana Trench in the western Pacific. In other words, there is no pre-200 Ma ocean floor remaining in modern oceanic domains, simply because non-stop oceanic subduction processes have kept consuming older ocean floors along active continental margins as regulated by continuous plate tectonics. Hence, we cannot retrieve any deep-sea record prior to the Early Jurassic from modern deep-sea floors, and one of the popular approaches in late Mesozoic–Cenozoic paleoceanography like deep-sea drilling is not applicable to studies of pre-Jurassic events including the P-TB event. On the contrary to such unfriendly conditions for us geologists, however, crucial oceanic subduction processes somehow did pay back benefits in the form of an accretionary complex (AC).

Ancient ACs include numerous remnants of pre-existing ocean floor material, such as deep-sea sediments, paleo-seamount/ogenetic plateau basalts with capping reef/atalt carbonates, and sometimes ophiolites from mid-oceanic ridge (Isozaki et al., 1990; see upper right diagram of Fig. 3) as typically observed in Japan. Due to their subduction–accumulation-related tectonic history, these rocks occur as allochthonous (exotic) blocks/lenses within younger matrices of trench-fill turbidites, and they are usually deformed, dismembered, and metamorphosed. Nonetheless, reconstruction of primary stratigraphy of mid-oceanic sediments is possible by utilizing high-resolution microfossil dating (e.g. conodonts,
radiolarians, fusulines etc.) as first demonstrated in Japan (e.g. Kanmera and Nishi, 1983; Matsuda and Isozaki, 1991). Such ancient accreted mid-oceanic sediments provide an important source of information for reconstructing paleo-oceanography and global environments not only of the extinction-related G-LB and P-TB cases (e.g. Isozaki, 1994, 1997) but also of the pre-Jurassic (=pre-200 Ma) world, as recently demonstrated in the Upper Neoproterozoic to Cambrian rocks by Uchio et al. (2004, 2008), Ota et al. (2007), and Kawai et al. (2008).

Under the unique paleogeographic configuration in the Permian characterized by a pair of single supercontinent Pangea and super-ocean Panthalassa, the ambient oceanographic systems at that time may have been considerably different from those of today. In particular, Panthalassa occupied nearly 70% of the Earth's

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**Fig. 1.** The Phanerozoic biodiversity change punctuated by major mass extinction events (A, redrawn from Sepkoski, 1984) and the Permo-Triassic details (B, redrawn from Knoll et al., 1996; Isozaki, 1997). The greatest mass extinction in the Phanerozoic comprises two distinct mass extinction events; i.e., one at the Guadalupian–Lopingian boundary (G-LB) and the other at the Permian–Triassic boundary (P-TB). Particularly noteworthy is (1) the sharp decline of marine animals with low metabolic rates (mostly sessile benthos or the Paleozoic fauna by Sepkoski, 1984 as shown in (A) with respect to those with high metabolic rates (animals with gills and strong internal circulation systems; i.e. Modern fauna by Sepkoski) at the G-LB, and (2) the greater magnitude of the selectivity at the G-LB than at the P-TB. Also note the low biodiversity interval from the G-LB to the mid-Anisian (Middle Triassic) for nearly 20 million years by and large overlaps the superanoxic period in the deep-sea (Isozaki, 1997).
were deposited in low-latitudes within 10 degrees of the paleo-data established that the lower Middle Triassic deep-sea cherts transport and subduction–accretion. In addition, paleomagnetic for the primary sedimentary site, prior to the long-term horizontal age plate convergence rate of about 3 cm per year, a ca. 3000 km Middle Jurassic accretion at the trench. Given a conservative average between the Late Permian sedimentation in the mid-ocean and the Jurassic AC in Japan, a ca. 100 million year-long age gap exists between the Late Permian AC in the Mino-Tanba belt (plus its lateral equivalents in the Chichibu belt and North Kitakami–Oshima belt) in Japan contain numerous allochthonous blocks of oceanic rocks (Isozaki et al., 1990). Among these, several large blocks of ancient mid-oceanic sediments retain primary stratigraphic sequences including both the P-TB and the G-LB intervals. Critical datasets for Permo-Triassic studies are two-fold; one from mid-oceanic deep-sea cherts and the other from mid-oceanic shallow-sea carbonates (Fig. 3), as described below.

In general, the primary deposition of these accreted oceanic rocks is difficult to locate but we can estimate their approximate positions in the superocean on the basis of some plate tectonic domain in the southern hemisphere (Kirschvink and Isozaki, 2007). Preliminary paleomagnetic measurements for the Middle–Upper Permian paleo-atoll limestone also suggest its origin in a low-latitude (12 degree) domain in the southern hemisphere (Kirschvink and Isozaki, 2007). During the Middle Permian to Early Triassic, therefore, the relevant oceanfloor was likely located somewhere in the equatorial Panthalassa.

In the following two sections, the essentials of bio- and chemostatigraphical aspects of deep-sea chert and atoll limestone are summarized separately.

3. Deep-sea chert

The stratigraphy of Late Permian to Early Triassic deep-sea chert in Japan, primarily derived from a mid-oceanic, deep-water domain of low-latitude Panthalassa has been analyzed in many sections during the last two decades (e.g. Yamakita, 1987; Isozaki, 1994; Kuwahara et al., 1998; Ezaki and Yao, 2000). The ubiquitous occurrence of framboidal pyrite and the relevant oceanfloor was likely located somewhere in the equatorial Panthalassa.

In the following two sections, the essentials of bio- and chemostatigraphical aspects of deep-sea chert and atoll limestone are summarized separately.

Fig. 2. Long-term changes in C and Sr isotope ratios and in sea level during the Phanerozoic (compiled from Veizer et al., 1999; Saltzman, 2005; Katz et al., 2005; McArthur and Howarth, 2004; Miller et al., 2005; Haq and Schutter, 2008). Note the G-LB as a unique timing with unusual geological signatures; i.e., the onset of volatile fluctuation in C isotope, and the Panthalassa minimum in Sr/Sr ratio and in sea level. The overall patterns of these three aspects do not appear in harmony with each other throughout the Phanerozoic; however, note that the G-LB marks a unique timing, at which these three independent phenomena synchronized to record the onset of unusual events of global scale.

The Mesozoic ACs in Japan and elsewhere in the circum-Pacific (Miyashiro-type) orogenic belts contain allochthonous blocks and lenses of Permo-Triassic mid-oceanic deep-sea cherts and paleo-atoll carbonates. The Late Permian AC in the Akiyoshi belt and the Jurassic AC in the Mino-Tanba belt (plus its lateral equivalents in the Chichibu belt and North Kitakami–Oshima belt) in Japan contain numerous allochthonous blocks of oceanic rocks (Isozaki et al., 1990). Among these, several large blocks of ancient mid-oceanic sediments retain primary stratigraphic sequences including both the P-TB and the G-LB intervals. Critical datasets for Permo-Triassic studies are two-fold; one from mid-oceanic deep-sea cherts and the other from mid-oceanic shallow-sea carbonates (Fig. 3), as described below.

In general, the primary deposition of these accreted oceanic rocks is difficult to locate but we can estimate their approximate positions in the superocean on the basis of some plate tectonic constraints. For example, in the case of those contained in the Jurassic AC in Japan, a ca. 100 million year-long age gap exists between the Late Permian sedimentation in the mid-ocean and the Middle Jurassic accretion at the trench. Given a conservative average plate convergence rate of about 3 cm per year, a ca. 3000 km separation from the eastern margin of the South China is assumed for the primary sedimentary site, prior to the long-term horizontal transport and subduction–accretion. In addition, paleomagnetic data established that the lower Middle Triassic deep-sea cherts were deposited in low-latitudes within 10 degrees of the paleo-equator (Shibuya and Sasajima, 1986; Oda and Suzuki, 2000; red star in the upper left map of Fig. 3). Preliminary paleomagnetic measurements for the Middle–Upper Permian paleo-atoll limestone also suggest its origin in a low-latitude (12 degree) domain in the southern hemisphere (Kirschvink and Isozaki, 2007). During the Middle Permian to Early Triassic, therefore,
3.1. Double-phased turnover of radiolarians

The Paleozoic radiolarians became almost completely extinct at the P-TB. Two major orders of Paleozoic radiolarians, Albaillellaria and Latentifistularia, disappeared at the end of the Changhsingian (Late Lopingian), whereas the Mesozoic-type orders, Nassellaria and Spumellaria, started to radiate in the Smithian, Early Triassic (De Weber et al., 2003). This major break in radiolarian lineages suggests development of an environmental stress of global scale at the P-TB, although Takemura et al. (2007) recently demonstrated a delayed radiolarian extinction in the high-latitude southern hemisphere. Because radiolarians, in particular the Paleozoic ones before the diatom dominance in the Mesozoic–Cenozoic world, represent the major marine plankton that regulated the

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**Fig. 3.** Composite stratigraphy and inter-facies correlation of the Middle Permian to Middle Triassic mid-oceanic deep-sea chert and shallow marine atoll carbonate in Japan that originated from the mid-superocean Panthalassa (modified from Isozaki, 2007b). The red star in the upper left map indicates the primary depositional site in low-latitude mid-Panthalassa of the mid-oceanic rocks. Upper right: a simplified diagram showing a ridge-trench transect showing the primary plate tectonic setting of deep-sea chert and paleo-atoll carbonates in mid-ocean (Panthalassa) before the subduction–accretion along the Pangean margins (modified from Isozaki et al., 1990). Columns are not to scale. Note the double-phased biotic turnover at the G-LB and the P-TB was recorded both in deep and shallow mid-Panthalassan sediments but in different manners.

<table>
<thead>
<tr>
<th>Age</th>
<th>Triassic</th>
<th>Permain</th>
<th>Guadalupian</th>
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<tr>
<td></td>
<td>Early</td>
<td>Lopingian</td>
<td>Capitanian</td>
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<td>Middle</td>
<td>Changhsingian</td>
<td>Wuchiagensian</td>
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<td>G-LB 260Ma</td>
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<td>P-TB 252Ma</td>
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<tr>
<td>Radiolarian</td>
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<td>N. ornithoformis</td>
<td>F. monacanthus</td>
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<td>Zone</td>
<td></td>
<td>N. optima</td>
<td>F. charveti</td>
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<td>T. deweveri</td>
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<td>F. ventricosus</td>
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<td>T. coronata</td>
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<td></td>
<td>F. scholasticus</td>
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<tr>
<td>H. gifuensis</td>
<td></td>
<td></td>
<td>Pseudoalbaillella globosa</td>
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<tr>
<td>P. nakatsugawaensis</td>
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**Fig. 3 continued:**

- Light gray limestone
- Red chert
- Gray claystone
- Black claystone

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Conodont/Fusiline Zone

P. bulgarica
C. timorensis
N. homeri
N. conservatius
N. waageni
N. dieneri
N. carinata
I. isarcica
H. parvus

Palaeofusulina
Codonofusiella-Reichelina
Lepidolina
Yabeina
Neoschwagerina
main silica factory in the ocean at the time, their collapse may indicate a considerable disturbance of the oceanic food web and a decline in marine productivity at various depths of the water column in the superocean.

Not much attention has previously been paid to; however, there was another radiolarian turnover among the order Albaillellaria immediately before the G-LB, in particular from the Middle Permian genus *Pseudoalbaillella* to the Late Permian *Neoalbaillella* via a transient form *Follicucullus* (Isozaki, 2007a; Fig. 4).

*Pseudoalbaillella* is a long-lived genus that diversified and constantly predominated in the Late Carboniferous to Middle Permian and they declined in diversity, in average body size, and also in population size around the G-LB (e.g. Ishiga, 1990; Yao and Kuwahara, 1999). The genus *Follicucullus* appeared in the Capitanian immediately before the G-LB. According to the biostratigraphy in West Texas and South China (Ormiston and Babcock, 1979; Nestell et al., 2006; Sun and Xia, 2006), the *Follicucullus* scholasticus-*F. ventricosus* (radiolarian) Zone, together with the underlying *F. monocanthus* Zone and the overlying *F. charveti* Zone, is correlated with the Capitanian. The precise G-LB horizon is still difficult to locate by radiolarians; however, it is tentatively placed at the top of *F. charveti* Zone.

In the Capitanian *F. monocanthus* Zone, *Pseudoalbaillella* became relatively smaller in body size and less abundant, whereas *Follicucullus* became dominant. In turn, *Follicucullus* became less dominant in the latest Capitanian (upper part of the *F. charveti* Zone). This quick turnover of Albaillellaria genera in the Capitanian is remarkable in contrast to their long-term stable lineage throughout the Late Carboniferous to Middle Permian, suggesting that the radiolarians faced a fatal change in environment immediately before the G-LB. The monotonous dominance of *Albaillella* plus *Neoalbaillella* throughout the Lopingian suggests that the oceanic condition was relatively stable up to the final extinction immediately before the P-TB.

3.2. Onset and climax of superanoxia

Another interesting observation about deep-sea chert is the development of a unique oceanographic condition called superanoxia across the P-TB; i.e., an extraordinarily long-term oxygen-depleted condition in the superocean Panthalassa (Isozaki, 1994, 1997). The pyritiferous dark-colored rocks across the P-TB in Japan and British Columbia represent deposition under considerably reduced up to euxinic conditions, in remarkable contrast with the Lower–Middle Permian and the Middle–Upper Triassic hematite-enriched brick-red cherts that accumulated under oxygenated conditions. The development of deep-sea anoxia across the P-TB has been suggested by various geochemical proxies, such as sulfur isotope ratios (Kajiwara et al., 1994), oxidation state of iron (Kubo et al., 1996; Matsuo et al., 2003), organic geochemistry (Suzuki et al., 1998), and the rare earth element (REE) abundance pattern (Ishiga et al., 1996; Kato et al., 2002) of the boundary sequence. In addition, the unusual occurrence of deep-sea dolomite nodules in the dark-colored boundary unit in Japan (Kondo and Adachi, 1975) also suggests that diagenetic precipitation of impure carbonates may have taken place under reduced condition even below...
changes appeared in two steps in Permian Panthalassa. In particular, a potential causal link is suggested between the redox change in the deep-sea and biodiversity change (Isozaki, 1997, 2007b). The apparent coincidence in timing between the mid-oceanic radiolarian turnover and the onset of the superanoxia supports the link. Along with the environmental changes in surface and moderately deep oceans, the Permian radiolarians were probably forced to change their physiology thus their test morphology specifically during this interval of redox change.

4. Paleo-atoll limestone

Numerous allochthonous blocks of Late Carboniferous to Triassic carbonates occur in the Late Permian to Jurassic ACs of Japan. These carbonates are mainly composed of shallow-marine bioclastic limestone (packstone/wackestone) with minor amounts of framstone, and they completely lack coarse-grained terrigenous clastics. They yield abundant shallow marine fossils of a typical Tethyan assemblage that suggests tropical to subtropical warm-water environments. Their basement consists of pillowed basaltic greenstones of oceanic-island (OIB)-type geochemical composition (e.g. Nishimura et al., 1979; Tatsumi et al., 2000). These lithologic characteristics and pre-accretion primary assemblage suggest that these limestones were deposited as a carbonate buildup or atoll complex on top of ancient mid-oceanic seamount probably of hotspot origin (Kanmera and Nishi, 1983; Sano and Kammera, 1988; see inset of Fig. 3). Preliminary paleomagnetic data suggest that the paleo-atoll limestone was deposited at 12° in southern hemisphere (Kirschvink and Isozaki, 2007). Some of these limestone blocks are of kilometric size and contain both P-TB and G-LB intervals (Koike, 1996; Isozaki and Ota, 2001).

Lithofacies, fusulin/conodont biostratigraphy, and stable carbon isotope signatures were analyzed for Middle Permian to Lower Triassic rocks in such accreted mid-oceanic limestone (Koike, 1996; Sano and Nakashima, 1997; Musashi et al., 2001, 2007; Ota and Isozaki, 2006; Isozaki et al., 2007a,b; Horacek et al., 2009; Fig. 5). In all these three aspects, changes occur also in a double-phased manner both at the G-LB and P-TB. Lithology of paleo-atoll carbonates changed gradually across the G-LB from the Guadalupian dark-gray limestone with TOC ~0.1 wt% to the Lopingian light gray limestone with lower TOC ~0.01 wt% (Ota and Isozaki, 2006). The top of the Lopingian limestone was dolomitized, and is directly covered by the Lower Triassic black microbial limestone (Sano and Nakashima, 1997). The P-TB-defining index conodont Hindeodus parvus was reported from ca. 50 cm above the base of this anachronistic microbialite (Koike, 1996).

4.1. Double-phased extinction of fusulines

The extinction of the Permian biota in a mid-oceanic shallow marine setting is best documented by fusulines (e.g. Kammera and Nakazawa, 1974; Sano and Nakashima, 1997; Ota and Isozaki, 2006), as the occurrence of conodonts/ammonoids is extremely rare in the Permian parts of the accreted Permo-Triassic limestone blocks due to facies control. Among the five Permian fusulin families, Schwagerinidae and Verbeekinidae first became extinct at the G-LB, whereas the rest three, Ozawainellidae, Schubertiidae, and Staffellidae, disappeared completely at the P-TB (e.g. Kammera et al., 1976; Ross, 1995). This double-phased extinction pattern of major fusuline families was examined not only in the mid-oceanic paleo-atoll limestone in Japan (e.g. Kammera and Nakazawa, 1974; Ozawa and Nishiwaki, 1992; Zaw Win, 1999; Ota and Isozaki, 2006; Fig. 5) but also in the coeval epeiric carbonate platforms around Pangea (Jin et al., 1994; Stanley and Yang, 1994; Wilde, 2002; Yang et al., 2004; Fig. 1B).
As to the G-LB extinction, a sharp faunal turnover was recognized between the Capitanian (upper Guadalupian) dark-gray limestone and the overlying Wuchiapingian (lower Lopingian) light gray dolomitic limestone in the Kamura section in Kyushu and the Akasaka section in central Japan (Sakagami, 1980; Ozawa and Nishiwaki, 1992; Zaw Win, 1999; Ota and Isozaki, 2006; Fig. 5). The Capitanian limestone is enriched solely in large-shelled verbeekinid fusulines, such as Yabeina and Lepidolina (up to 1 cm in diameter), whereas the Wuchiapingian rocks lack large individuals and contain only much smaller-shelled (less than 2 mm in diameter) schubertellids, ozawainellids, and staffellids, such as Codonofusisella, Reichelina, and Staffella. The Wuchiapingian interval is dominated by Reichelina and Codonofusisella, whereas the Changhsingian part by Palaeofusisella, Staffella, and Nankinella (Kammera and Nakazawa, 1974; Sano and Nakashima, 1997; Ota and Isozaki, 2006). All these Lopingian fusulines as a whole were terminated consequently at the P-TB as examined at two separated sections in Japan, i.e., Kamura and Taho (150 km apart) (Koike, 1996).

The Kamura, Taho, and Akasaka sections in Japan, where the double-phased extinction of Permian fusulines was detected, are currently separated from each other by ca. 300–500 km. The identical extinction pattern confirmed at the physiographically separated multiple sections implies that the extinction and relevant environmental changes in two steps were not local but regional phenomena in low-latitude Panthalassa, and probably of global context. Such a double-phased extinction pattern of fusulines suggests that significant environmental changes affecting the shallow-water benthic protists may have appeared in the surface of mid-Panthalassa at least twice.

4.2. Double-phased negative shift of C-isotope

Chemozonotigraphical analyses utilizing stable carbon isotopes have also detected a double-phased change around the G-LB and the P-TB in the paleo-atoll limestones at Kamura in Japan (Fig. 5). First across the G-LB, the secular change in carbonate carbon isotope ratio ($\delta^{13}$C$_{\text{carb}}$) demonstrated a gradual drop of ca. 5‰ around the fusuline-defined G-LB (Isozaki et al., 2007a). The underlying Capitanian interval is characterized by extremely high positive $\delta^{13}$C$_{\text{carb}}$ values of over +5‰ up to +7‰, whereas the overlying Wuchiapingian by relatively low positive ones of +2 to +3‰. As such high positive values between +5 and +7‰ are unusual throughout the Permian and unique solely to this interval, this phenomenon was named the Kamura event that may represent a high-productivity event in mid-Panthalassa, and probably a short-term cooling in the Capitanian immediately before the G-LB (Isozaki et al., 2007a,b). It is noteworthy that the total negative shift occurred after the main extinction and spanned the biostratigraphical G-LB defined by the first occurrence of the Lopingian taxa. The negative shift around the G-LB corresponds to that at the GSSP of G-LB at Penglaitan, South China (Wang et al., 2004). Unfortunately, $\delta^{13}$C$_{\text{org}}$ values at Kamura are still under measurement, and the results will be reported elsewhere shortly.

The other prominent negative shift of stable carbon isotope ratio was recognized at the P-TB in two separate sections in Kamura and Taho (Musashi et al., 2001, 2007). A clear negative spike occurred both in carbonate carbon ($\delta^{13}$C$_{\text{carb}}$) and organic carbon ($\delta^{13}$C$_{\text{org}}$) (Fig. 5). Both $\delta^{13}$C$_{\text{carb}}$ and $\delta^{13}$C$_{\text{org}}$ values decrease for ca. 2‰, showing a clear parallelism between them that likely suggests universal $^{13}$C-depletion in ambient seawater. This negative shift has been detected in many sections in the world and these sections were mutually correlated (e.g., Holser et al., 1989; Baud et al., 1989).

Although the Lopingian interval has not yet been thoroughly documented, the C isotope chemostratigraphic study on the accreted Permian limestone clarified an overall decreasing trend from the Late Middle Permian to the earliest Triassic punctuated by double-phased sharp negative shifts at the G-LB and the P-TB. This trend is in general agreement with that compiled by Korte et al. (2005). As the G-LB and the P-TB are separated from each other by ca. 300–500 km, the double-phased negative shift of C-isotope might reflect a global environmental change affecting the shallow-water benthic protists.
other by ca. 8 million years, the double-phased isotopic shift requires two independent driving mechanisms to cause a negative shift in C isotopic composition of seawater twice on a regional scale in the superocean.

4.3 Sr minimum

The secular change in strontium isotopes in the Guadalupian to early Lopingian interval was recently demonstrated in mid-Panthalassan paleo-atoll carbonates at Kamura (Kani et al., 2008). With respect to the $^{87}\text{Sr}/^{86}\text{Sr}$ values of modern seawater and of many Paleozoic carbonates in equilibrium with paleo-seawater (e.g. Veizer et al., 1999; Fig. 2), the values of the Kamura samples are concentrated obviously in a much lower value range. They show a remarkable change in a long-term trend from a gradual decrease in the Wordian–Capitanian to an increase in the Capitanian–Wuchiapingian. The minimum value 0.706914 ± 0.000012 was recorded in the Yabeina Zone (lower–middle Capitanian) at Kamura (Fig. 5). This Capitanian minimum in Sr isotope ratio detected in mid-Panthalassa atoll carbonates readily corresponds to that detected in coeval carbonates of the peri-Pangean shelf domains (e.g., Veizer et al., 1999; Korte et al., 2006), suggesting a global signature. Marking a chemostratigraphically unique turning point in Panthalassic trend, this minimum indicates that a remarkable change occurred in the Capitanian global oceanography with respect to the Sr-isotope balance of seawater between mantle- and continental fluxes. In order to explain this unique event, Isozaki (2007b) and Kani et al. (2008) suggested that the initial rifting of Pangea might have been responsible in providing a huge amount of continent-derived (radiogenic Sr-enriched) terrigenous clastics into the Panthalassan Ocean on the basis of the occurrences of LIPs in the eastern Pangea. Further $^{87}\text{Sr}/^{86}\text{Sr}$ analysis is under way with respect to the P-TB interval.

5. Discussion

Possible causes and processes of the end-Permian double-phased event are explored here on the basis of the above-introduced new lines of evidence from the two contrasting mid-oceanic sedimentary facies of Panthalassa, i.e., deep-sea chert and shallow marine atoll carbonates in addition to the data from the peri-Pangean domains. The discussion starts from the geological implications of the double-phased extinction mainly from the Panthalassan perspective, moves onto the interpretation of isotope records of carbon and strontium, and finally deals with possible causal mechanisms with particular emphasis on mantle plume activity and related environmental changes on the surface.

5.1 Double-phased biotic turnover/extinction

As to the double-phased extinction, two distinct animal groups hold the key roles in the food web of the extensive mid-superoceanic domains, i.e., radiolarians and fusulines. In the Late Paleozoic, radiolarians were the most dominant plankton in open water, whereas fusulines represent one of the major benthos in shallow marine environments. Their double-phased extinction pattern confirmed in the mid-Panthalassa, concordant with that in the peri-Pangean continental margins, suggests that major environmental stress(es) of global scale occurred twice in the Middle–Late Permian biosphere. Both radiolarians and fusulines likely formed relatively basal parts of the Permian marine food web, therefore, the rise and fall in their diversity and in abundance probably recorded globally-averaged environmental states including nutrient availability in moderately deep to very shallow levels of the superocean.

5.1.1. Radiolarian turnover in the deep superocean

Radiolarians are marine plankton whose habitable zone ranges from surface water down to more than a thousand meter depth in modern oceans. The Late Paleozoic radiolarians probably occupied by and large the same habitat in the ocean as modern examples. In the extensive Panthalassan domain, Permian radiolarians suffered double-phased faunal turnover; i.e., significant faunal change around the G-LB and final extinction at the P-TB. As to the G-LB turnover, *Pseudoalabaillella*, the long-ranged and long-dominant genus since the Late Carboniferous declined in abundance, in diversity, and also in average body size for the first time in the Permian (Ishiga, 1990; Kuwahara et al., 1998). The lineage was consequently taken over by the Lopingian genus *Neoalabaillella*, via a transient genus *Follisculius* that first appeared in the Guadalupian. This likely suggests that a certain kind of unfavorable situation for the long-lasting radiolairian taxa, and that the relevant environmental stress may have emerged in mid-Panthalassa and forced the radiolarian group to change physiology in terms of test morphology by the end of the Guadalupian. Possible triggers include the oxygen-depletion in seawater, to which the radiolarian test structure may have adjusted in regard of metabolism; e.g., efficiency in gas exchange, in nutrient uptake, and/or in waste-dump through more perforated test. It is noteworthy that the radiolarian turnover started not right at G-LB time but slightly earlier in the early Capitanian (Fig. 4). Nonetheless, the long-term deposition of radiolarian chert per se was not much affected, if at all, because the Milankovitch-tuned, rhythmically-bedded chert accumulated continuously all the way from the Guadalupian almost to the late Lopingian. Consequently the total radiolarian production did not change significantly across the G-LB, despite the lineage shift.

In contrast, the P-TB event was critical for the Paleozoic radiolarians. Except for rare survivors in the southern hemisphere (Takemura et al., 2007), all the Paleozoic forms became extinct in the topmost Changoisian (e.g., Yao and Kuwahara, 1999; De Weber et al., 2003). The P-TB interval of the deep-sea facies in Japan, from the topmost Changoisian to the lower Olenekian (Smithian, Triassic), is composed of radiolarian-free (or -poor) siliceous claystone/black shale without any chert (chert gap; Fig. 4). The deposition of the unique P-TB black shale in the deep-sea may have resulted from unusual bacterial blooming after the radiolarian extinction. The earliest recovery of radiolarians after the P-TB event was in the Dienerian in western Tethys (Kozur, 1996). In other words, the radiolaria-driven Permian silica factory in the mid-ocean collapsed almost completely in the latest Changoisian, and became dormant during the Griesbachian (ca. 1 million years immediately after the P-TB; Fig. 4). The abrupt disappearance of radiolarians at the P-TB and their retarded recovery were confirmed in the contemporary shallow marine continental shelf facies in South China (e.g., He et al., 2005; Jin et al., 2007; Isozaki et al., 2007c). The direct kill mechanism has not yet been identified; however, their total extinction possibly reflects a major collapse in marine productivity on a global scale because the Paleozoic radiolarians had a wide geographical distribution from low-latitudes even to higher latitudes, probably more extensive than modern ones due to the absence of diatoms in the Paleozoic. The abrupt extinction of radiolarians also suggests that a large perturbation appeared not only in surface waters but also in the moderately deep-water of mid-Panthalassa. Although the slightly delayed extinction in the high-latitude southern hemisphere was reported (Takemura et al., 2007), this does not change the general trend of radiolarian transition from the Paleozoic to the Mesozoic-type.
After all, radiolarians, the representative Paleozoic marine plankton of generally cosmopolitan nature, suffered major faunal turnover twice in the second half of the Permian. The impact of the P-TB event was obviously more profound than that of the G-LB. This indicates that the long-lasting normal food web of the Late Paleozoic abruptly collapsed in the mid-superocean at the P-TB. Nonetheless, from a long-term viewpoint, it is noteworthy that the major change in the Permian plankton started not at the P-TB but around the G-LB (Fig. 4), and in particular, that the initial change started not at the biostratigraphically-defined G-LB but sometime earlier in the Capitanian. For making comparison with the G-LB extinction in the deep superocean, coeval extinction in shallow mid-Panthalassa is discussed next.

5.1.2. Extinction in the shallow superocean

Fusulines represent a shallow-marine, warm-water, benthic foraminifera group that probably thrived within the euphotic zone (less than 50 m deep) just like modern benthic foraminifera. They successfully flourished mainly in low-latitude areas throughout the Late Carboniferous–Permian not only along the peri-Pangean continental margins but also on isolated mid-Panthalassan paleo-atolls, and they declined in diversity in two steps by the end of the Permian (e.g., Jin et al., 1994; Stanley and Yang, 1994; Ota and Isozaki, 2006). The first fusuline extinction at the G-LB resulted in a sharp reduction in test size, suggesting the appearance of an environmental stress to screen out selectively large-tested taxa. This phenomenon may correspond to the so-called “post-extinction Lilliput effect” in a broad sense (e.g., Urbanek, 1993; Twitchett, 1999). According to the classic ecological observations on modern biota, a dominance of small-sized individuals called r-strategists generally indicates the appearance of a stressful condition in environments. The “Lilliput effect in fusulines” appeared at the G-LB, and continued for another 8 million years throughout the Lopingian until the total extinction of all fusulines was completed at the P-TB (Fig. 5). This suggests that the Late Permian shallow marine environments and relevant ecological structure were essentially different from the long-lasting stable ones during the Early-Middle Permian.

The Early-Middle Permian fusuline faunas are generally dominated by large-, robust-shelled taxa, i.e., Schwagerinidae and/or Verbeekinidae (Fig. 5). In particular, the late Guadalupian fusuline fauna in low-latitude realms of Tethys, Panthalassa, and western North America, consist of extraordinarily large-tested genera (e.g., Yabeina, Lepadolina, and Polydidioidina) up to 1 cm in diameter, with minor amounts of small ones (e.g., Kammer et al., 1976; Ishii, 1990; Sheng, 1990; Ross, 1995). Permian large-tested fusulines evolved to form highly complicated wall structures including keriotheca, whereas the smaller ones were conservative and retained simple and thin (primitive) wall structures. In analogy with modern foraminifers, the complicated skeletal structures with many volutions and sophisticated wall fabrics in large fusulines likely represent an adaptation for hosting symbiotic algae and/or cyanobacteria (Ross, 1972; Wilde, 2002; Vachard et al., 2004). In addition to fusulines, some Permian gigantic bivalves, corals, and brachiopods in shallow tropical seas are also listed as fossil photosymbiotic animals (Cowen, 1983; Seilacher, 1990; Isozaki, 2006), and their diversity change along time is also interesting as discussed below.

The Permian giant fusulines became extinct abruptly at the G-LB, both in mid-oceanic paleo-atoll limestones (Ota and Isozaki, 2006) and in continental shelf carbonates (e.g., Sheng, 1990; Stanley and Yang, 1994; Ross, 1995; Wilde, 2002; Yang et al., 2004). In contrast, the lower Wuchiapingian rocks yield only simpler-structured and smaller fusulines less than 2 mm in diameter, e.g. Staffiello and Nankinella. Wilde (2002), Yang et al. (2004), and Ota and Isozaki (2006) proposed that the sharp screening of the large fusulines at the end of the Guadalupian might have been related to the dying-out of associated symbiotic algae. It is noteworthy that the rise and fall of large fusuline diversity occurred almost at the same time with those of the Permian gigantic bivalves and various rugose corals (Wang and Sugiyama, 2002; Isozaki and Aljinovic, in press). Paying special attention to the gigantism in shallow marine invertebrates in tropical oligotrophic environments, Isozaki and Aljinovic (in press) further emphasized the synchronous extinction of the photosymbiosis-dependent “tropical trio” (large fusulines, large bivalves, and rugose corals) in shallow Tethys and Panthalassa. A large-scale environmental change may have occurred in shallow marine environments at the end of the Guadalupian preferentially in the photic zone of low-latitude Tethys and Panthalassa. Possible direct kill mechanisms may include temperature drop, decreased insolation, change in nutrient level, and/or salinity change; however, details are not yet clarified. Isozaki and Aljinovic (in press) proposed that the Capitanian cooling (Kamura event; Isozaki et al., 2007a,b) coupled with eutrophication might have terminated the photosymbiotic fauna that were highly adapted to warm-water, oligotrophic conditions.

On the basis of the fossil data from the peri-Pangean shelf facies, Knoll et al. (1996) pointed out the selective screening by the G-LB extinction of animals with no gills, weak internal circulation and low metabolic rate, i.e., mostly sessile benthos with passive respiratory systems, such as cnidarians, echinoderms, and articulate brachiopods. Powers and Bottjer (2007, 2009) added supporting evidence from the on/off-shore migration pattern of the Permian brachiopods. The recent analysis on the extinction pattern of various taxa across the G-LB (Clapham et al., 2009) suggests a rather gradual turnover during the second half of the Guadalupian rather than a rapid extinction specifically at the G-LB.

The common occurrence of the barren interval in the topmost Capitanian paleo-atoll sections (Ota and Isozaki, 2006) is noteworthy because this confirms the common sequence of stages in ecological transition relevant to extinction; (1) appearance of environmental stress, (2) main extinction, (3) prolonged survival, and (4) recovery. The kill mechanism obviously started to operate already in the late Capitanian at the latest, and this agrees also with the radiolarian extinction pattern in the mid-ocean (Fig. 4), as discussed above. Like other major extinction cases in the Phanerozoic, the extinction of the Guadalupian shallow marine fauna started before the biostratigraphic G-LB that is defined by the first occurrence of the Lopingian fauna (Fig. 5). The environmental stress since the late Capitanian (e.g., temperature drop) likely suppressed the biodiversity and population size for certain duration until the environment was ameliorated in the Early Wuchiapingian.

At the P-TB, on the other hand, various shallow marine biota, not only fusulines and rugose corals but also others (smaller foraminifers, brachiopods, gastropods, crinoids, and calcareous algae) suffered severely both in peri-Pangean shelf environments and in mid-Panthalassan paleo-atolls. As in the case of the G-LB extinction, the biotic turnover and responsible environmental change were of global extent; however, the kill mechanism for the P-TB case may have been slightly different from that for the G-LB. As a classic explanation, the appearance of short-term cooling was proposed as a possible cause for the mass killing at the end of the Permian (Stanley, 1988; Kozur, 1998). In addition, other possible kill mechanisms of shallow marine fauna were proposed on the same track that essentially assumes an upward migration of chemocline in the superocean; such as suffocation by oxygen depletion (anoxia), hypercapnia by elevated CO2, and poisoning by H2S (e.g., Wignall and Hallam, 1992; Knoll et al., 1996; Riccardi et al., 2006).

In the beginning of the Early Triassic that is claimed as an interval of retarded recovery, the characteristic anachronistic
microbialites and unique Lilliput fauna occur both in peri-Pangean continental shelves and in mid-Panthalassan paleo-atolls (Sano and Nakashima, 1997; Kershaw et al., 1999; Lehrmann, 1999; Woods et al., 1999; Twitchett, 1999; Ezaki et al., 2003; Fraiser and Bottjer, 2004; Pruess et al., 2005), indicating that the post-P-TB aftermath may have experienced a short-term return to the Perminian world characterized by shallow marine communities of reduced paleoecological tiering under prolonged environmental stresses. The delayed recovery after the P-TB event took much longer than that after the G-LB event.

5.2. Double-phased carbon isotope shift

In addition to the fossil records, the drastic environmental change in the Permian–Triassic mid-superocean is clearly recorded in some geochemical proxies, such as chemical state of iron, REE abundance pattern, and isotope ratios of C, S, and Sr. The secular change in stable carbon isotope ratio monitored an unusually volatile fluctuation of the global carbon cycle during the Paleozoic–Mesozoic transition interval (Figs. 2 and 4). In particular, the double-phased nature of the secular change is noteworthy. These data confirmed not only that the Middle Permian to Lower Triassic mid-Panthalassan limestones are bio- and chemographically correlated with the peri-Pangean/Tethyan shelf carbonates, but also that the Late Permian ocean seawater has run into an isotopically extreme condition twice across the G-LB and the P-TB in global context. These two episodic events in the global C cycle are discussed here one by one.

5.2.1. G-LB episode

A significant negative shift in δ13C_carb for ca. 5‰ was recently detected across the G-LB horizon in a mid-oceanic paleo-atoll limestone (Isozaki et al., 2007a; Fig. 5). This shift started appreciably after the extinction of the Guadalupian large-testest fusulines and before the first appearance of the Lopingian fauna. As being correlated with δ13C_carb stratigraphy of peri-Pangean/Tethyan shelf facies at the GSSP of the G-LB at Laibin, South China (Wang et al., 2004), this negative shift across the G-LB likely has a global context. There are three interesting aspects to be noted in the C isotope stratigraphy across the G-LB; i.e., (1) the negative excursion of δ13C_carb in mid-Panthalassa occurred in three steps, interspersed with two steps of short-term positive shifts, (2) the total amount of negative shift across the G-LB reached ca. 5‰, much greater than that of the P-TB (ca. 2‰), and (3) the shift occurred after the main extinction event. A negative shift of δ13C of organic carbon across the G-LB was detected also at the Penglaitan GSSP (Kaiho et al., 2005), and the essentially parallel negative shift between δ13C_carb and δ13C_org values indicates that the biological pump using the same C-fixation processes (most likely that of oxygenic photosynthesis) was still operating in shallow mid-Panthalassa even after the G-LB extinction. As detailed δ13C_org data have not yet been available from deep-sea chert except for preliminary ones (Ishiga et al., 1993), the present author and his colleagues are working on this problem, and the results will be reported elsewhere.

More emphasis should be given to the Capitanian interval with unusually high positive δ13C_carb values around +5‰ to +6‰ (Fig. 5) rather than to the negative shift in δ13C_carb across the G-LB. The high δ13C_carb interval apparently lasted for ca. 3 million years, that has never been reported from the Permian, and this unique isotopic episode named Kamura event likely records a cooling period in the Capitanian may explain this. The Middle Permian has been generally regarded as a post-glaciation warming period after the Carboniferous to Early Permian Gondwana icehouse; however, the appearance of a cold snap in the late Middle-Late Permian is suggested by some geological observations in high-latitudes (e.g., Uistritsky, 1973; Beauchamp and Baud, 2002; Fielding et al., 2008). The extinction of the “tropical trio” in low-latitudes (Isozaki, 2006; Aljinović et al., 2008; Isozaki and Aljinović, in press) and the first migration of the Permian mid-latitude brachiopods into the tropical domain (Shen and Shi, 2002) during the latest Guadalupian likewise support this interpretation, though the global influence of the claimed Kamura cooling event should be further tested elsewhere.

In contrast, the above-mentioned G-LB negative shift per se in the aftermath of the Kamura event, likely represents the onset of global warming. During the interval of a negative excursion in δ13C_carb but still with positive values, the Lopingian fauna appeared, and the biodiversity started a post-extinction recovery (Ota and Isozaki, 2006).

5.2.2. P-TB perturbation

In the mid-oceanic paleo-atoll carbonates spanning the P-TB, the secular change both in δ13C_carb and δ13C_org values recorded a clear negative shift (Musashi et al., 2001, 2007; Fig. 5). This negative excursion in δ13C_carb is well correlated with that in many sections of marine and terrestrial facies around the world (e.g., Holser et al., 1989; Baud et al., 1989; Wang et al., 1994; Twitchett et al., 2001) including the GSSP at Meishan in South China (Jin et al., 2000b; Yin et al., 2001).

The overall parallelism between δ13C_carb and δ13C_org in the Upper Permian–Lower Triassic shallow marine carbonates in mid-Panthalassa and also in Pangean margins (Magaritz et al., 1992; Musashi et al., 2001, 2007; Fig. 5) proves (1) that dissolved inorganic carbon in shallow-sea water globally became depleted in 13C across the P-TB, and on the other hand, (2) that photosynthesis-driven isotopic fractionation was still working generally on the surface even after the extinction-related P-TB event. In the Griesbachian immediately after the extinction of various Permian algae and metazoans associated with the collapse of normal ecological structures, bacteria (including photosynthetic cyanobacteria) may have been dominant in C-fixation to keep the marine productivity, as evidenced by the anachronistic (Precambrian-like) microbialities in both Pangean shelves and mid-Panthalassa atolls (e.g., Schubert and Bottjer, 1992; Sano and Nakashima, 1997; Kershaw et al., 1999; Lehrmann, 1999).

In contrast, however, the P-TB deep-sea black shale recorded a positive spike in δ13C_carb, showing the opposite trend to the coeval signature in shallow marine carbonates (Fig. 5). The Lopingian and Dnienerian–Anisian deep-sea claystone/cherts have δ13C_org values around ~30‰ ( Ishiga et al., 1993; Wada et al., unpublished data), whereas the contemporaneous shallow marine carbonates on paleo-atolls record δ13C_org values around -25‰ (Musashi et al., 2001). On the other hand, the Griesbachian black shale uniquely recorded δ13C_org values around ~27‰, and the contemporaneous shallow marine carbonates also have the same values around ~27‰. This illustrates that the Late Permian to Middle Triassic oceans generally had a depth-wise isotopic gradient in Corg of ca. 5‰ between the shallow and deep water (Fig. 6 upper); however, this gradient disappeared during the Griesbachian (Fig. 6 lower).

The development of such a remarkable isotopic gradient with depth in the Permo-Triassic Panthalassa except the Griesbachian interval was likely driven by selective decomposition of organic matter in the oxygenated upper water column of the ocean. Among the components of organic matter, proteins (amino acids) are more easily decomposed by bacterial activities under oxygenated conditions than lipids, in particular, during their sinking as particles through subphotic zone in the upper water column (e.g., Wakeham...
As lipids usually have relatively light C-isotope ratios with respect to proteins, the preferential decay of proteins results in a relative abundance of lipids in organic matter deposited in deep-sea sediments that likely drives lower $\delta^{13}$C$_{org}$ values. Under the ordinary oceanic conditions of the (partly or fully) ventilated Late Paleozoic–Mesozoic deep-sea, therefore, a clear gradient developed in $\delta^{13}$C$_{org}$ values between shallow and deep waters. As long as protein-screening in the oxic upper water column worked to maintain the same isotopic gradient, any change in the carbon source in the surface ocean may likewise have been recorded in the C-isotope of deep-sea organic matter; i.e., the negative shift of $\delta^{13}$C$_{org}$ for 2‰ across the P-TB in shallow marine carbonates (solid green line in the inset of Fig. 6) should also have been recorded in deep-sea black shale (broken green line). This indicates that the clear $\delta^{13}$C$_{org}$ gradient of 5‰ persisted during the Changhsingian eventually disappeared across the P-TB, and such an unusual monotonous condition continued at least during the Griesbachian. The absence of this gradient in the Griesbachian indicates that the preferential decomposition of protein was suppressed in the water column, likely by the upward excursion of the chemocline almost to the surface (lower). Consequently the unique oceanic condition (superanoxia) enhanced the direct downward transport and burial of organic carbon in deep-sea.

In remarkable contrast, however, the Griesbachian records show that the shallow atoll carbonates and deep-sea black shale have more or less the same $\delta^{13}$C$_{org}$ values (solid red and green lines in the inset of Fig. 6), in other words, the depth-wise gradient in the C-isotope of deep-sea organic matter; i.e., the negative shift of $\delta^{13}$C$_{org}$ for 2‰ across the P-TB in shallow marine carbonates (solid green line in the inset of Fig. 6) should also have been recorded in deep-sea black shale (broken green line). This corresponds to this case with more effective transportation of proteins to the deep-sea. In the Lopingian, deep-seated anoxic water increased in volume to push the chemocline upward, then around the P-TB at the zenith of superanoxia, the chemocline may have reached to a very shallow level of ocean (Isozaki, 1994) as confirmed by the biomarker for photic zone euxinia (Grice et al., 2005; Hays et al., 2007). The unique “superanoxic condition” in the Griesbachian superocean likely drove an unusual, isotopically monotonous oceanic structure all the way from bottom to top (Fig. 6 lower), and it ended when the chemocline retreated back to the deep-sea; i.e., re-oxygenation of deep ocean.

5.3. Possible causes for the double-phased extinction

The above-mentioned double-phased environmental change associated with major extinction, in particular, the secular changes of C and Sr isotopes clearly identified the G-LB and P-TB events as two independent events, highlighting the long-overlooked significance of the G-LB event and the uniqueness of the Late Permian interval. The G-LB and P-TB events were clearly separated for ca. 8 m.y., therefore, these two independent events should have been triggered by two distinct causes of global influence. Given that everything occurred in a double-phased manner, all the geological phenomena during the second half of the Permian can be summarized in the framework of two sets of cause, process, and result. As
to the possible causes of the biggest mass extinction in the Phanerozoic, however, previous discussion naturally focused mostly on the P-TB event.

The most popular scenario at present for the P-TB event is volcanism-induced environmental turmoil and resultant extinction (e.g., Wignall, 2002; Erwin, 2006; Bottjer et al., 2008), because the apparent coincidence in timing between extinction and large-scale volcanism, in particular continental flood basalt (CFB) such as the Siberian Traps, has often been emphasized (e.g. Campbell et al., 1992; Renne et al., 1995; Courtillot, 1999). Several varieties of extinction scenario were derived from this assumption of the CFB-extinction link; e.g., the emission of voluminous volcanic gas into the atmosphere and the associated melting of methane hydrates to release more CO₂ for an apocalyptic super-greenhouse condition in the biosphere (Retallack, 1999; Wignall, 2002; Huey and Ward, 2005; Meyer and Kump, 2008). Possible direct kill mechanisms include (1) suffocation by O₂ depletion (Wignall and Hallam, 1992), (2) hypercapnia with excess CO₂ (Knoll et al., 1996), and (3) poisoning by H₂S (Riccardi et al., 2006). All these hypotheses assume volcanism-triggered global warming and associated oceanic stratification on a global scale. In accordance with the P-TB superanoxia, the upward migration of chemoline in the world ocean is commonly speculated. For generating such a large-scale change in ocean structure of global extent, the temperature gradient between equatorial and polar regions, likely played an important role (e.g. Hotinski et al., 2001). Nonetheless the onset mechanism of such a unique oceanographic condition remained enigmatic, and this will be discussed later. The hypothesis of P-TB global warming induced by CFB volcanism/methane hydrates may explain the delayed recovery of biotic diversity during the Early Triassic but not necessarily the cause of the extinction at the P-TB per se. As to the timing, however, there is still an appreciable time gap between the extinction and CFB volcanism. Furthermore, the extinction slightly but apparently predated the CFB volcanism. More precise dating is definitely needed for checking the potential cause–effect link between the CFB volcanism and extinction. An alternative scenario with volcanism-related kill mechanism emphasizes the importance of felsic magmatism associated with continental rifting (“plume winter” scenario by Isozaki (2007b)), and it is explained in the next section.

On the other hand, by applying the K-T boundary scenario, some have claimed an extraterrestrial bolide impact as the cause of the P-TB event (e.g., Becker et al., 2001; Kairo et al., 2001); however, failures in reproducing identical geochemical and mineralogical signals from the same sample and also in identifying the exact boundary horizon have hampered broad acceptance of the idea (e.g., Farley et al., 2005; Isozaki, 2001; Koebel et al., 2002). The last claim for a possible contemporary impact crater off Western Australia (Becker et al., 2004) was also refuted for various reasons (e.g. Renne et al., 2004). In fact the impact crater in Australia is inconsistent with the fact that the contemporary off-shore of New Zealand, located not far away from Australia, served as the sole exceptional refugia for the Paleozoic radiolarians during the P-TB extinction event (Takemura et al., 2007). The apparent chronological similarity may suggest a possible causal link between extinction and extraterrestrial impact (or CFB volcanism) but cannot prove it without solid lines of material-based evidence, as we experienced in the K-T boundary controversy until the discovery of the culprit crater in Yucatan.

The cause of the G-LB event has been discussed less frequently with respect to the P-TB. The volcanism of the Emeishan Traps in western South China and the Panjal Traps in northern India was likewise claimed as a possible cause of the G-LB extinction on the basis of their chronological coincidence (e.g., Chung et al., 1998; Courtillot, 1999; Zhou et al., 2002; Ali et al., 2002; He et al., 2007). However, the nominated CFBs in S. China and India are apparently too young and too small to have been responsible for all the above-listed strange geologic phenomena that started within the Capitanian. Just like the P-TB case, the significance of felsic magmatism needs more attention also for the G-LB case (Isozaki and Ota, 2001; Isozaki et al., 2004), as will be discussed in the next section. There is no evidence at all for an extraterrestrial impact around the G-LB, thus we need not pursue the idea of impact of large bolides.

As mentioned above, the ultimate causes of the G-LB and P-TB events have not been identified yet; however, the current data indicate that no bolide impact scenario is necessary for both cases, instead that terrestrial processes, such as formation of LIPs by the mantle plume activity, were somehow related. The G-LB and P-TB events share some similarities, but on the other hand, they also have clear disparities. Thus it is obvious that the two events do not represent the consequences of simple repetition of the same cause and processes. In the following section, at first, common trigger and processes between the two events are discussed with special emphasis on non-basaltic magmatism. Then in the second next section, differences between the two events are reconsidered with particular focus on the unique geomagnetic phenomenon that appeared in the late Guadalupian, and a possible scenario is explored for total explanation of all similarities and disparities between the two extinction-relevant events.

5.4. Double-phased “plume winter”

On the basis of all the available datasets from both shallow marine and deep-sea facies in Panthalassa and Tethys, Isozaki (2007b) proposed the “plume winter” scenario, in order to explain the greatest turmoil of the Phanerozoic biosphere that took place in the transition interval from the Paleozoic to the Mesozoic-modern world. This hypothesis emphasized that the greatest mass extinction was led by extraordinarily large-scale magmatism probably derived from a mantle superplume impinged beneath the supercontinent Pangea, therefore, at a glance, it may appear similar to the volcanism-related scenarios previously proposed (e.g., Campbell et al., 1992; Renne et al., 1995; Chung et al., 1998; Courtillot, 1999). The plume winter scenario, however, focuses not on CFB as suggested by many but on plume-generated felsic alkaline volcanism, simply because tuff beds around the G-LB and P-TB are all felsic in composition.

In searching for the ultimate cause among various major events during the Paleozoic-Mesozoic transition interval, the most important is likely the felsic volcanism that occurred uniquely around the G-LB and P-TB. The concentrated occurrence of felsic volcanism at these timings suggests their origin not from steady-state magma sources but from episodic ones. The boundary felsic tuffs were recognized extensively in many P-TB sections in South China (e.g., Yin et al., 1993). It is noteworthy that the onset of felsic volcanism slightly predated the timing of final extinction, thus such volcanism appears critical in assessing the causal mechanism of the extinction-relevant environmental changes. For example, more than 10 tuff beds across the P-TB (within a 4 m-thick interval spanning the uppermost Changhsingian and lowermost Griesbachian) were confirmed in northern Sichuan (Li et al., 1989; Isozaki et al., 2004, 2007c). One of these tuffs in the middle horizon is precisely dated at 251–252 Ma as the P-TB marker bed in South China (e.g., Bowring et al., 1998; Mundil et al., 2004; U-Pb zircon age by ID-TIMS, not by SHRIMP).

The G-LB tuff in South China has long been regarded as a claystone under the traditional name “Wangpo bed” (Lu, 1956; Li et al., 1991); however, this unit represents a bona fide air-borne tuff that contains abundant volcanogenic material including euhedral phenocrysts of quartz, plagioclase, apatite, and zircon (Isozaki et al., 2001, 2004, 2008; Isozaki, 2007b). The Wangpo tuff, ca.
1–2 m thick, occurs throughout western South China from Shaanxi to Guanxi, covering nearly 1000 km in an N–S direction (Isozaki et al., 2004) and this G-LB tuff in northern Sichuan has a U-Pb zircon SHRIMP age of 260 ± 4 Ma (He et al., 2007). In addition, the occurrence of the Capitanian fine-grained tuff beds in mid-oceanic sequences (Isozaki, 2007b) suggests much more extensive delivery range of ash; i.e., Western Panthalassa together with South China was covered on a regional scale by air-borne tuff of rho-yo-dacitic composition in the Late Guadalupian. An unusually large-scale and high explosive volcanism of the Plinian to Ultratplinian-type can solely explain such a long-distance delivery and a huge volume of tuffaceous material. In contrast to mafic, felsic volcanism usually occurs with a violent, explosive eruption that can explain the putative environmental catastrophe that may have driven extinction.

The occurrence of the prominent Permian LIPs and contemporary volcanicogenic sediments (Isozaki, 2007b) indicates that large-scale volcanism of both felsic and mafic composition occurred in the second half of the Permian, probably peaked twice around the G-LB and the P-TB. Steady-state plate boundary processes along mid-oceanic ridges and arc-trench systems continue all the time as long as plate tectonics is in operation on the Earth. Thus relevant volcanisms keep active without having episodic or catastrophic pulses that can drive unusual events like mass extinction. Instead, within-plume volcanism derived from mantle plumes likely had a high episodicity in forming LIPs (Ernst et al., 2007). In fact, some LIPs of the Permian age accompany anadestic basalt and rhyolite units, e.g. the Emeishan and Siberian Traps; however, these felsic units occur limitedly in the middle and upper parts of the volcanic edifice, thus were too young to have caused the extinction-related environmental change. They are also too small in amount to have accounted for the entire coverage of the boundary tuffs, as mentioned above. As to the Emeishan Traps, He et al. (2007) speculated for a voluminous rhyolites above the main basalt unit and their extensive erosion on the basis of geochemistry of the overlying Upper Permian rocks: however, the generation of claimed large amount of felsic magma from the source rocks (mostly mantle peridotite), with respect to the volume of basaltic one, is unlikely because of multiple steps of partial melting required for producing felsic magma. The non-basaltic composition of the Upper Permian sediments probably reflects the erosion of sialic (continental) crust of older basement or pre-basalt felsic volcanics.

By considering the above issues, Isozaki (2007b) speculated that an episodic activity of mantle superplume might have driven the G-LB and P-TB events, because the Paleozioc–Mesozoic transition corresponds to the time of initial rifting of Pangea prior to its main breakup in the Mesozoic. As to the breakup of a supercontinent, the uprising of a superplume and its impingement at the continental lithosphere are vital. At some time in the Guadalupian, a superplume was likely launched from the deep mantle, particularly from the D"{o}"{o} layer (ca. 2600–2900 km deep) immediately above the core-mantle boundary in order to counterbalance in mass for the descending cold plume or megalith (subducted oceanic slabs) (Maruyama, 1994; Maruyama et al., 2007).

A mantle superplume likely branches into several secondary plumes in the mantle transition zone (ca. 410–660 km deep) where the density contrast between a plume and surrounding mantle decreases remarkably according to the mineral phase transition (Maruyama, 1994; Maruyama et al., 2007). The Middle Permian superplume probably branched into plural secondary plumes, which impinged at the bottom of the supercontinent in two steps to cause a “plume winter” twice; i.e. around the G-LB and P-TB.

The impingement of a main plume head beneath Pangea probably generated strong magmatism by decompressional melting of mantle peridotite (+ recycled slabs). Prior to the main flood basalt eruption, however, re-melting of pre-existing continental sialic crust likely occurred to generate felsic magmatism. This may reasonably explain the order of the two distinct types of volcanisms at the G-LB and P-TB; i.e. first felsic then mafic, not vice versa. In addition, eruption of gaseous kimberlite may have been accompanied as a precursor of plume magmatism (Morgan et al., 2004; Isozaki, 2007b), in particular, when a superplume impinged at a CO2-enriched tectosphere beneath relatively older continents (Santosh et al., 2009). Ultratplinian-type eruptions of these felsic plus kimberlitic volcanoes at multiple sites in swarm may have driven severe environmental changes in the biosphere through volcanic hazards; e.g. toxic gas emissions, developing dust/aerosol screens in the stratosphere (blocking sunlight), and pouring acid rain. Their cascading effects on the environments, such as temperature drop, dim daylight, cessation of photosynthesis, and shortage in food, may have led to the decline in biodiversity both in the ocean and on land; i.e. a condition called “plume winter” (Isozaki, 2007b). Although actual kill mechanisms should be scrutinized more in detail for each animal group both in marine and terrestrial settings, all the responsible mechanisms likely became effective by the global development of the background “plume winter” conditions.

The first arrival of the Permian superplume at the bottom of Pangea was reflected in the Phanerozoic minimum in the 87Sr/86Sr ratio in the Capitanian marine sediments (Kani et al., 2008) that is well correlated with peri-Pangean shelf carbonates (Veizer et al., 1999; Korte et al., 2006; Fig. 2). Since the Cambrian, 87Sr/86Sr ratios of marine carbonates kept decreasing for nearly 300 million years regardless of minor fluctuations; however, they suddenly changed the trend in the Late Guadalupian and started to increase rapidly toward the early Mesozoic high, suggesting that a large-scale reorganization had occurred in the Sr isotope balance in the Panthalassic seawater between the mantle flux and the continental flux. In general, low sea levels induce active erosion of continental crusts that increases 87Sr/86Sr ratio of seawater. During the Capitanian with the lowest sea level in the Phanerozoic, however, the 87Sr/86Sr ratios of marine carbonates globally marked the lowest values. Isozaki (2007b) and Kani et al. (2008) interpreted that this contradictory behavior of seawater Sr isotope ratio have been related to the initial rifting of Pangea in the late Guadalupian. By making new drainage systems directly connected the intra-Pangean basins and Panthalassa, the rifting may have triggered abrupt mass wasting of terrigenous clastics that accumulated within Pangea since its formation in the Late Paleozoic.

As to the P-TB, no measurements on Sr isotopes have been made of mid-oceanic rocks; however, the previous data from continental shelf carbonates demonstrate that 87Sr/86Sr ratios uni-directionally increased across the P-TB (Veizer et al., 1999) merely with a short-term standstill (Twitchett, 2007). This highlights the uniqueness of the trend change and abrupt increase of Sr-isotope around the G-LB, and proves that no major erosion/mass wasting occurred in Pangea across the P-TB. In addition, it is noteworthy that the Sr-isotope fluctuation in seawater was essentially independent from that of the stable carbon isotope around the P-TB interval, although both C and Sr isotope systems started to shift drastically at the same time not at the P-TB but in the Capitanian.

It is emphasized here that both the G-LB and P-TB extinctions were probably led by plume-generated felsic/kimberlitic volcanism, because flood basalt volcanism took place slightly later than the main extinction in both cases. The relatively lower viscosity of basalt magma cannot cause explosive eruption sufficient for delivering air-borne tuff over great distances. Instead, the input of a huge amount of CO2 to the atmosphere by plume-driven volcanism (all of kimberlitic, felsic and mafic) may explain the post-extinction apocalyptic global warming (a plume summer) coupled with a sea-level rise and the retarded post-extinction recovery in the biosphere in the Early Triassic (Twitchett, 1999). In short, a pair of “plume winter” and “plume summer” appeared twice around the
G-LB and P-TB probably by the two-stepped activity of branched secondary plumes.

The mechanism of developing the 20 m.y.-long superanoxia is still enigmatic. In general, a stratified ocean is difficult to maintain for millions of years under the oxygenated atmosphere, as long as the planet Earth keeps its rotation and temperature gradient exists between the polar and equatorial regions, as demonstrated in the numerical modeling (e.g. Hotinski et al., 2001). The apparent conundrum of the long-term development of deep-sea anoxia, however, may be explained reasonably by assuming two steps of oxygen depletion caused by two independent triggers (Isozaki, 2007b). The earlier half of the superanoxia (during the early Lopingian) was probably related to the first plume summer that suppressed the ocean mixing after the G-LB. The anoxia may have been gradually relaxed with fluctuation in redox during the late Lopingian. Then, before the superocean returned to a fully oxic condition, the second plume summer (by the Siberian Traps) appeared to drive the anoxia into an ultimate state across the P-TB (the zenith of superanoxia), and it took long time to be cancelled until the early Middle Triassic. The initial trigger was not related to the plume volcanism.

6. Integrated “plume winter” scenario: volcanic cooling and geomagnetic cooling

As discussed above, the G-LB and P-TB events share many common geological characteristics, suggesting that the two events have similar causes probably related to mantle superplume activity. Nonetheless, clear differences exist between the two events, indicating that not exactly the same cause and processes repeated. By referring to the unique geologic phenomena that occurred particularly in the late Guadalupian, I here present an integrated version of the “plume winter” scenario at the end of this article.

Fig. 7 illustrates the entire chronicle of the peculiar changes in the core, mantle, crust, and surface biosphere of the planet during the Paleozoic–Mesozoic transition interval that was punctuated twice by the G-LB and P-TB events. The G-LB event, or more precisely speaking, the Capitanian event should be much emphasized simply because everything geologically unusual appeared at first in the Capitanian. In fact, after the long quiescence since the mid-Carboniferous Pangean assembly and ca. 8 million years before the P-TB per se, various unique changes on the Earth’s surface started during the Capitanian (Isozaki, 2007b). In addition to the major faunal turnover (Jin et al., 1994; Stanley and Yang, 1994) and the onset of LIP-forming magmatism (Isozaki, 2007b), the following features uniquely characterize the Capitanian event; i.e., (1) onset of superanoxia (Isozaki, 1997), (2) onset of volatile C-isotope fluctuation (Isozaki et al., 2007a,b), (3) trend change in Sr-isotope ratio (Veizer et al., 1999; Kani et al., 2008), (4) sea-level change from a long-term regression to a transgression (Haq and Schutter, 2008), and (6) pattern change in geomagnetic polarity (Gradstein et al., 2004; Steiner, 2006).

The most unique feature in the above-listed geologic phenomena related to the G-LB event is the sharp pattern shift in geomagnetic polarity called the Illawarra Reversal (Fig. 8). This remarkable change in geomagnetic field reversal pattern took place around the
Wordian–Capitanian boundary (ca. 265 Ma), marking a sharp transition from the long-lasting Kiaman Reverse Superchron (Late Carboniferous to Middle Permian) to the Permian–Triassic Mixed Superchron (Late Permian to Late Triassic) (e.g., Jin et al., 2000; Gradstein et al., 2004; Steiner, 2006). The Illawarra Reversal was recently confirmed also in the mid-Panthalassan paleo-atoll carbonates in Japan (Kirschvink and Isozaki, 2007; Fig. 5). Such a remarkable change in the behavior of the Earth's magnetic field provides direct evidence for the appearance of a major thermal instability at the core-mantle boundary that may have been related to the probably synchronous launching of a superplume (Isozaki, 2009). When the stability of Earth's dipole field declined, a decrease in geomagnetic intensity could have been triggered by alteration of molten iron circulation patterns in the outer core, driven in turn by events at the core-mantle boundary (e.g., Tarduno et al., 2006; Courtillot and Olson, 2007).
Through consideration of the superplume-launching event in the late Middle Permian, Isozaki (2009) further extended the "plume winter" scenario to connect the unique event in history of the geodynamo and geomagnetic field with the concurrent

**Fig. 9.** Schematic diagrams of the integrated "plume winter" scenario. The episodic activity of a mantle superplume beneath the supercontinent Pangea was the most likely cause of the double-phased mass extinction at the G-LB and P-TB. (A) The Illawarra Reversal and the Kamura event (modified from Isozaki, 2009): Around the beginning of the Capitanian (late Middle Permian) (ca. 265 Ma), a dramatic change took place at the core/mantle boundary (D’ layer) by the drop of a relatively cold megalith (subducted oceanic slabs beneath Pangea) from the mantle transition zone (410–660 km deep). The prominent geomagnetic event called Illawarra Reversal (Fig. 8) recorded the appearance of thermal instability on the core’s surface that drove a major mode change of the geodynamo in the outer core. The unstable dipole likely lowered geomagnetic intensity of the field, thus increased the flux of galactic cosmic rays to the atmosphere. By the enhanced cloud coverage, the Earth’s albedo increased to trigger the Capitanian Kamura cooling event (Figs. 5 and 7) that resulted in the selective decline/extinction of the unique tropical fauna and the unusually high oceanic productivity coupled with the expansion of O2 minimum zone in mid-depth. It is noteworthy that the sea level reached the lowest of the Phanerozoic and the superanoxia started around this time. (B) A plume winter condition (modified from Isozaki, 2007b): The breakup of the supercontinent Pangea was intimately linked to the impingement of a superplume-head at the base of a supercontinental lithosphere nearly 5 million years after the Illawarra Reversal, i.e., around the G-LB (ca. 260 Ma). In the mantle transition zone (ca. 410–660 km deep), a mantle superplume branched into plural secondary plumes due to the density contrast of rocks related to mineral phase transitions along depth. During the 20 million years of the Paleozoic–Mesozoic transition, two episodes of "plume winter" occurred intermittently, i.e., at the G-LB and P-TB. These were triggered by two sets of secondary plumes that branched from the main superplume in two steps. Through the violent magmatism of felsic and partly kimberlitic nature, various volcanic hazards (volcanic cooling, toxic gas emission, pouring acid rain, and sunlight-blocking dust/aerosols) led the Earth’s surface to a "plume winter" condition. Besides everything else, the shutdown of sunlight and global cooling by the dust/aerosol screen (plume winter) appear to have been the most significant because the cessation of photosynthesis on a global scale is critical in maintaining bio-diversity. A short-time darkness and coldness can halt photosynthesis on a global scale and disorganize the pre-existing food web. The following flood basalt volcanism drove the surface environment into global warming (plume summer), delayed post-extinction recovery in biodiversity, and intensified the stratification of the superocean Panthalassa. Although caused by the similar superplume-related processes, the G-LB event involved both geomagnetic cooling and volcanic cooling, whereas the P-TB event lacked the former process. The lowest biodiversity was recorded across the P-TB event probably because the time interval was too short between the two episodes for retrieving the full biodiversity back to the Paleozoic plateau level (Fig. 1A).
changes in the biosphere around the G-LB (Fig. 9A). For launching a huge mass from the deep mantle, the fall-down of a mass of subducted slabs (megalith) is necessary from the upper mantle to the D* layer above the core-mantle boundary, and this provides the most compelling mechanism for the destabilization of the geomagnetic dipole. Such an event would place a huge, relatively cool mass directly onto the bottom of mantle, driving both upwelling of a warm superplume through volume displacement and rapid propagation of thermal instabilities from the D* layer into the outer core. In fact, this is the one and only practical way to modify the mode of geodynamo rapidly and drastically. Thus the Illawarra Reversal recorded in surface rocks provides a likely prime marker for the launching of a superplume from the core-mantle boundary around 265 Ma.

Destabilization of the geomagnetic dipole may increase the likelihood of frequent polarity changes and may also weaken field intensity, as often observed in short-term fluctuations in the Quaternary. Assuming a constant galactic cosmic ray flux, increasing cosmic ray penetration and bombardment of the Earth's atmosphere would track with decreasing geomagnetic field intensity as the Earth's field provides the primary armor deflecting away high-energy cosmic radiation (Tinsley, 2000). Although the fundamental processes of cloud formation are still under scrutiny (e.g., Haigh, 2007), an increase in high-energy cosmic ray flux may enhance both ionization of atmospheric molecules (aerosols) and resulting formation of cloud nuclei; i.e., the more cosmic radiation, the more cloud coverage over the Earth (e.g., Dickinson, 1975; Svensmark, 2007). Along this vein of reasoning, increased cloud cover leads to increased albedo, decreased sunlight (heat) flux to the Earth's surface, and decreased global surface temperatures.

If this were the case for the Illawarra Reversal in the Guadalupian, a geodynamo-driven decrease in surface temperature could reasonably explain the Capitanian Kamura cooling event coupled with the lowest sea level in the Phanerozoic (Figs. 2, 5, 7 and 8A; Isozaki, 2009). Global cooling may have accelerated ocean mixing, providing an increased nutrient supply to the surface ocean to drive the Kamura high-productivity event. Increased surface productivity, in turn, may have expanded the mid-depth oxygen-minimum zone of the oceans and knocked the oxic, early Capitanian deep-sea into anoxia. The onset of the superanoxia was likely related not to a global warming by plume-generated volcanism but to the Capitanian Kamura cool event, whereas the developed deep-sea anoxia was maintained for a while under the following global warming across the G-LB. In this regard, the drawdown of atmospheric CO₂ recorded in the δ¹³Ccarb was not a likely cause of the cooling as originally proposed (Isozaki et al., 2007a) but probably one of the consequences of the geomagnetic cooling.

Around 260 Ma, nearly 5 m.y. after the putative superplume launching from the core-mantle boundary, the first magnetism derived from secondary plumes appeared on the surface, leading to the first “plume winter” (Fig. 9B). This event likely accelerated global cooling to complete the G-LB extinction. Against the currently popular interpretations of global warming and relevant extinction both at the G-LB and P-TB, the significance of cooling driven by mantle plume activity is emphasized in this article. The two extinction-bearing boundary events likely shared “volcanic cooling” induced by secondary plumes; however, the P-TB event did not experience “geomagnetic cooling”. This made the major difference between the two events and highlighted the uniqueness of the G-LB event or the Capitanian interval.

At the end of discussion, other geologic periods with long-term stable geomagnetism are checked from the similar viewpoints (Fig. 8). The mid-Cretaceous Normal Superchron (ca. 118–83 Ma) demonstrates another such link between geomagnetic field stability and climate, but going in the opposite direction. The Earth’s atmosphere during the Long Normal Superchron, a geomagnetically stable interval, would have experienced a low cosmic ray flux and thus higher anticipated surface mean temperatures. This is in agreement with the long-term warm climate for the mid-Cretaceous coupled with a high sea level, thermally stagnant oceans, and resulting episodic oceanic anoxia (e.g., Larson, 1991). Detailed magnetostratigraphy of the Orдовician is still unknown; however, the “relatively stable” period called Burskan reversed bias polarity interval (particularly its later part; Algeo, 1996) or Moyero Reverse Superchron (ca. 490–460 Ma; Pavlov and Gallet, 2005) appears concordant with the long-lasting warm period that allowed the rapid diversification of metazoans (e.g., Servais et al., 2009).

In contrast, the mid-Carboniferous to mid-Permian Kiaman Reversal Superchron (ca. 310–265 Ma) chronologically corresponds to a much colder period, including the Gondwana glaciations (Gradstein et al., 2004). This case apparently contradicts the geomagnetic field stability/climate model. Shaviv and Veizer (2003) resolved this contradiction, however, by explaining two independent variables for controlling the cosmic ray flux, i.e., the intensity of radiation source(s) and the strength of magnetic shield by Sun and Earth. Cosmic ray flux into the Earth’s atmosphere depends not only on geomagnetic and helio-magnetic field intensity but also on the number, size, and proximity of galactic cosmic ray sources. Due to the proximity between our solar system and a supernova-ridden spiral arm of the Milky Way galaxy, the flux of cosmic radiation increased in the mid-Carboniferous to Early Permian interval (including the Kiaman Superchron), independent of changes in the geo- and helio-magnetic field. The period of this oscillation is estimated ca. 143 ± 10 m.y. (Shaviv and Veizer, 2003) that is much longer than that for the heliomagnetism and much shorter than that of superplume-induced geomagnetic rhythm, but concordant with the well-known long-term cycle of greenhouse-icehouse of the Phanerozoic. As to the two remarkable snowball Earth events in the Paleo- and Neo-proterozoic, i.e., the 2.3–2.2 Ga and 0.7–0.6 Ga events, Veizer (2005), Svensmark (2007), and Maruyama et al. (submitted) further speculated for the cosmo climatological link between the major cooling event in history and increased galactic cosmic radiation in terms of the distance from the neighboring galaxies.

7. Conclusion

Recent studies in Japan on the accreted ancient mid-oceanic rocks have provided vital information on the unique double-phased global event that occurred in the lost Panthalassa superocean during the Paleozoic–Mesozoic transition interval. In addition to the conventional data from the peri-Pangean shallow marine settings, all the new information from Panthalassa confirmed that the era-boundary event between the Paleozoic and Mesozoic was not a simple short-term episode, like the end-Cretaceous case with one large meteorite impact and a resultant mass extinction. Instead, the event was composed of much longer-ranged complicated geologic/geophysical phenomena that took place in various parts of the Earth from its core to the surface. In particular, the double-phased nature of the extinction and relevant environmental changes at the G-LB and P-TB was clearly demonstrated in a global context, and this requires at least two independent causes that were chronologically separated for at least 8 m.y. from each other. Evidence from field geology suggests that a superplume activity in mantle most likely played the main role in extinction-related environmental changes at the G-LB and P-TB. The double-phased felsic (plus kimberlitic) volcanism, double-phased environmental turmoil, and double-phased extinction (triple–double) during the Paleozoic–Mesozoic transition interval, are by and large summarized into two repeated sets of trigger, process, and consequence that were essentially driven by
double-phased activity of secondary plumes branched from a main superplume. The plume-derived violent volcanism in two steps likely caused “plume winter” twice at the G-LB and P-TB to drive the extinctions. The G-LB event (or Capitanian event) was distinct from the P-TB event, however, as associated with the lowest sea level, the lowest Sr-isotope ratio, onset of superanoxia, and the prominent change in geomagnetic stability.

The integrated version of the “plume winter” scenario appears promising because, in this interpretation, a number of geological phenomena occurred in multiple spheres of the Earth, from its core to surface, and uniquely in the Paleozoic-Mesozoic transition interval, and because it likely explains all concordantly in one picture. Many parts of this “grand-unified explanation” presented here, however, are obviously too immature to be validated immediately, therefore, challenging new interpretation needs to be checked and polished through serious criticisms from many quarters. Nevertheless, this explanation may revive general interests in seeking possible interactions between terrestrial and extraterrestrial phenomena, in a completely different way from the widely known bolide impact scenario. The viewpoint and methods used for analyzing the Paleozoic-Mesozoic transition interval may provide a template for further application to paleoenvironmental studies on the pre-200 Ma deep past, i.e., the Paleozoic and Precambrian time. In particular, re-evaluation from alternative perspectives is needed for major events in the history of biosphere and life evolution during the Precambrian.

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