Guadalupian–Lopingian boundary event in mid-Panthalassa: Correlation of accreted deep-sea chert and mid-oceanic atoll carbonate

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Abstract

The Middle–Late Permian or Guadalupian–Lopingian (G-L) boundary marks a major mass extinction event comparable in magnitude to the one at the Permo-Triassic (P-T) boundary, the greatest in the Phanerozoic. The Middle–Upper Permian mid-oceanic rocks derived from the lost superocean Panthalassa, including deep-sea chert and shallow marine atoll carbonates, occur as large allochthonous blocks in a Jurassic accretionary complex in Japan. These accreted oceanic rocks record vital information on the global environmental change at the G-L boundary. The dataset from the accreted oceanic rocks in Japan is two-fold, one related to mid-oceanic deep-sea environment and the other to mid-oceanic surface waters. The deep-sea chert in Japan records a remarkable faunal reorganization in radiolarians, a major marine plankton group in the Permian, at the G-L boundary. Also recognized around the G-L boundary horizon is the onset of a long-term deep-sea anoxia that lasted for ca. 20 million years from the Lopingian to Anisian across the P-T boundary. On the other hand, mid-oceanic shallow-sea carbonates primarily deposited on ancient seamounts record a remarkable turnover in fusulines, and a ca. 3 per mil negative shift in δ¹³C_carb from +5 down to +2 per mil. Both mid-oceanic deep- and shallow-water sequences indicate that a significant change occurred in oceanography of the superocean across the G-L boundary. A unique rhyo-dacitic tuff bed occurs at the G-L boundary both in deep-sea sediments and shallow-sea paleo-atoll carbonates. This G-L boundary felsic tuff is correlated with a coeval tuff bed in continental shelf carbonates in South China. This strongly suggests that extensive areas, including western Panthalassa and South China, were influenced by a severe felsic volcanism at the G-L boundary. The rhyo-dacitic geochemistry of the volcanic material indicates that the responsible eruption(s) may have been highly violent and thus explosive enough to trigger global scale environmental turmoil and relevant mass extinction at the G-L boundary.

Keywords: mass extinction, Panthalassa, Permo-Triassic boundary, paleo-seamount, deep-sea chert, felsic volcanism.

Introduction

The end-Permian was a time when Earth’s biosphere drastically changed from the Paleozoic to a modern regime. A huge proportion of the Paleozoic marine and terrestrial invertebrates was terminated almost instantaneously at the Permo-Triassic (P-T) boundary and Mesozoic pioneer taxa filled the empty ecospace in the aftermath (e.g., Sepkoski et al., 1984; Erwin, 1993). Despite nearly a century-long discussion over the ultimate cause of this extinction event, no general agreement has yet been reached (e.g., Hallam & Wignall, 1998; Erwin et al., 2002). The apparent coincidence in timing suggests that the extinction may have been caused by strong volcanism, such as responsible for the Siberian Traps (e.g., Campbell et al., 1992; Renne et al., 1995), but no material-based evidence has yet been given. An interpretation of extraterrestrial bolide impact challenges this deadlock when applied to the same scenario for the K-T boundary event (Becker et al., 2001; Kaiho et al., 2001). Failures in identifying the exact boundary horizon and in reproducing identical geochemical and mineralogical signals from the same sample have hampered acceptance of the impact model (e.g., Farley & Mukhopadhyay, 2001; Isozaki, 2001; Koeberl et al., 2002; Hender-
son & Wardlaw, 2003). The latest claim for a possible contemporary impact crater off Western Australia (Becker et al., 2004) is refuted because neighbouring New Zealand served as the sole exceptional refugia for the Paleozoic radiolarians across the P-T boundary extinction event (Takemura et al., 2003).

Since the detailed biostratigraphic analyses by Jin et al. (1994) and Stanley & Yang (1994), the end-Paleozoic decline in biodiversity has been regarded as two independent mass extinction events that occurred successively within a short time period; i.e., one at the Middle–Late Permian or Guadalupian–Lopingian (G-L) boundary and the other at the P-T boundary sensu stricto. The great magnitude of the end-Paleozoic massacre probably resulted from the close temporal separation between the two independent events that did not allow full recovery in biodiversity immediately after the first (Stanley & Yang, 1994).

Compared with the P-T boundary event, not much attention was paid to the G-L boundary event before 1994, probably owing to a lesser awareness amongst geologists. Nevertheless, the G-L boundary has profound implications quite distinct from those of the P-T boundary. First, the G-L boundary marks the initial major decline after the long-lasting biodiversity plateau during the Late Paleozoic since the end-Devonian, as clearly shown by Sepkoski (1984, 1996). In particular, the diversity loss at the G-L boundary is greater in non-mobile, benthic animals rather than in self-locotomous or free-swimming ones, suggesting that a harsh environment in the ocean may have appeared to selectively terminate immobile dwellers on the seafloor (Knoll et al., 1996; Fig. 1).

Second, the G-L boundary corresponds to the onset in timing of a long-term global deep-sea anoxia called the superanoxia (Isozaki, 1994). At the same time, extinction of radiolarians and fusulines occurred in a mid-oceanic realm (Isozaki, 1997a; Isozaki & Ota, 2001; Ota & Isozaki, 2006). These changes indicate that a major reorganization of the oceanographic regime had started at the G-L boundary in the superocean Panthalassa. Particularly noteworthy is the apparent temporal coincidence between the mass extinction and the onset of deep-sea anoxia that may suggest a possible cause–effect relationship.

This article reviews recent progress in G-L boundary research on ancient mid-oceanic rocks primarily derived from the superocean Panthalassa. All these mid-oceanic rocks are currently contained in the accretionary complexes in Japan. On the basis of the correlation between deep-sea chert and shallow-sea paleo-atoll limestone, I discuss here the G-L boundary extinction, relevant environmental change, and a possible cause–effect link between the mass extinction and felsic volcanism at the G-L boundary.

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**Fig. 1. Secular changes of the Early Permian to Middle Triassic biodiversity (modified from Knoll et al., 1996).** Note the two-stepped decline in biodiversity in the Late Permian, i.e., one at the Capitanian (Guadalupian)–Wuchiapingian (Lopingian) boundary (G-L boundary) and the other at the Changhsingian–Induan boundary (P-T boundary). Particularly noteworthy is the sharp decline of non-mobile benthos at the G-L boundary that is more significant than that at the P-T boundary. For the putative end-Guadalupian global environmental change, the Permian non-mobile benthos were likely less tolerant than free swimmers and locomotives. The low biodiversity interval from the G-L boundary to mid-Anisian (Middle Triassic) apparently corresponds to the superanoxic period in the deep-sea (Isozaki, 1994).
Records in accreted mid-oceanic sediments

Ancient accretionary complexes include remnants of pre-existing ocean floor materials, such as deep-sea sedimentary rocks, fragments of paleo-seamount/oceanic plateau with capping reef/atoll carbonates, and sometimes ophiolites from mid-oceanic ridge (Kanmura & Nishi, 1982; Isozaki et al., 1990; Fig. 2). Owing to their tectonic history in the Asian active margin, many of these rocks are often deformed, dismembered, and metamorphosed, however, reconstruction of primary stratigraphy of mid-oceanic sediments is possible by utilizing high-resolution microfossil dating (e.g., conodonts and radiolarians) as first introduced in Japan (Matsuda & Isozaki, 1991). Such ancient accreted mid-oceanic sediments are an important source of information for establishing pre-Jurassic oceanography and global environments because the oldest extant ocean floor is not older than 200 Ma (Early Jurassic), i.e., the Pacific floor along the Mariana Trench in the western Pacific. Under the unique end-Permain configuration of continents and oceans at this time, with Panthalassa occupying nearly 70% of the Earth’s surface, the ambient oceanographic system at that time may have been considerably different from that of today. Thus, for reconstructing the end-Permain (ca. 250 Ma) global environment and the double-phased mass extinction, an analysis of ancient accreted oceanic sediments, i.e., a source of information of past basins, is inevitable.

The Late Permian accretionary complex in the Akiyoshi Belt and the Jurassic accretionary complex in the Mino-Tanba Belt (including the tectonic outlier of a nappe called the Chichibu Belt on the Pacific side) in Southwest Japan contain numerous allochthonious blocks of oceanic rocks (Isozaki, 1997b, c). Among these, several large blocks of ancient mid-oceanic sediments retain primary stratigraphic sequences including both the P-T boundary and the G-L boundary intervals. Datasets are two-fold; one from mid-oceanic deep-sea cherts and the other from mid-oceanic shallow-sea carbonates (Fig. 2), as described below.

It is difficult to identify the precise location of primary deposition of these accreted rocks, but we can still estimate their approximate positions in the superocean. In the case of those contained in the Jurassic accretionary complex, for example, 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, ca. 3,000 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 the subduction-accretion (Isozaki, 1997a; Isozaki & Ota, 2001).

Deep-sea chert

A Late Permian to Early Triassic deep-sea chert in the Mino-Tanba belt primarily derived from a mid-oceanic, deep-water domain of low-latitude Panthalassa has been analyzed biostratigraphically for conodonts and radiolarians in many sections (refer to compilation and summary by Isozaki, 1996; Ezaki & Yao, 2001). Additional data, more or less in the same

Fig. 2. Simplified diagram showing a ridge-trench transect showing the primary plate tectonic setting of deep-sea chert and paleo-atoll carbonates (modified from Isozaki et al., 1990). Crustal rocks of an oceanic plate are transported horizontally to an arc-trench system by oceanic subduction, and the mid-oceanic sedimentary rocks are often peeled off tectonically from the main oceanic plate to be incorporated into an accretionary complex. Late Permian and Jurassic accretionary complexes in Japan were formed along the ancient South China margin, and contain numerous exotic blocks from mid-Panthalassa, e.g., deep-sea chert, paleoatoll limestone, and basaltic greenstone of hotspot origin. Not to scale.
context, were obtained from British Columbia and New Zealand (Isozaki, 1997a; Takemura et al., 2003). These chert beds mostly consist of radiolarian tests with a minor amount of clay minerals, as reflected in their high silica (ca. 95 wt%) and low aluminium (less than 3 wt%) contents, and coarse-grained terrigenous clastics are absent.

By virtue of their fine-grained nature and quartz-dominated simple mineralogy, the chert sequences in the Mino-Tanba belt record a unique paleo-redox episode during the Late Permian to Early Triassic interval in iron-bearing minerals (Isozaki, 1994, 1997a; Fig. 3). The dark colours of the interval rocks and ubiquitous occurrence of pyrite indicate deposition under an oxygen-depleted condition. The long-term oxygen-depletion is supported by various geochemical proxies, such as sulfur isotope ratios (Kajiwara et al., 1994), the oxidation state of iron (Nakao & Isozaki, 1993; Kubo et al., 1996; Matsuo et al., 2003), organic geochemistry (Suzuki et al., 1998), and the rare earth element (REE) abundance pattern (Kato et al., 2002). The rocks immediately beneath and above the P-T boundary, in particular, may have been deposited under considerably reduced (dysoxic and anoxic) and partly euxinic conditions. These dark-coloured rocks across the P-T boundary show a remarkable contrast with the Lower–Middle Permian and the Middle–Upper Triassic brick-red cherts that accumulated under oxygenated conditions. Almost identical redox histories detected in Japan and British Columbia support...
the global nature (at least in the northern hemisphere) of this deep-sea anoxia named superanoxia (Isozaki, 1994) that persisted for nearly 20 million years across the P-T boundary. A putative short-term oxygenation in deep-sea environments at the P-T boundary during a long-lasting anoxia interval (Kajiwara et al., 1994; Suzuki et al., 1998) is kept out of consideration here, following the criticism of Wignall & Twitchett (2002).

During this superanoxic interval in the superocean, all the Paleozoic-type radiolarians apparently became extinct around the P-T boundary (Isozaki, 1994; Kozur, 1998; Ezaki & Yao, 2001). As radiolarians represent the major marine plankton that served as the main factory of biogenic silica in the ocean at the time, their collapse may indicate a considerable decline in marine productivity at various depths of the water column in the superocean. 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 P-T boundary when the superanoxic event reached its zenith, although Takemura et al. (2003) recently pointed out a delayed radiolarian extinction in the Middle Permian well-oxygenated red chert conformably overlain by a dysoxic dark-colored one of Late Permian age as confirmed in the Sasayama and Gujo-Hachiman sections in Japan (Ishiga et al., 1982; Kuwahara, 1999). The color transition occurs immediately below the F. scholastics–F. ventricosus (radiolarian) Zone. Therefore, the redox transition from oxic to anoxic, i.e. onset of the superanoxia, is regarded to have occurred around the G-L boundary.

### Paleo-atoll limestone

Late Carboniferous to Triassic carbonates of paleo-atoll origin often occur as kilometric-scale allochthonous blocks in Late Permian and Jurassic accretionary complexes of Japan. These carbonates are mainly composed of shallow-marine bioclastic limestone (packstone/wackestone) with minor amounts of framestone but completely lack coarse-grained terrigenous clastics. Their basement consists of basaltic greenstones of oceanic-island (OIB)-type geochemical composition. This relationship suggests that the overlying carbonates were deposited on top of ancient mid-oceanic seamounts probably of hotspot origin (e.g., Nishimura et al., 1979; Tatsumi et al., 2000) as carbonate buildups or as atoll complexes (Kanmera & Nishi, 1982; Isozaki, 1987; Fig. 2).

The P-T boundary interval in such limestones was analyzed from various aspects including fusuline plus conodont biostratigraphy, lithofacies analysis, and stable carbon isotope measurements. In the Kamura limestone in Kyushu, a sharp faunal/lithologic change occurs immediately below the conodont-defined P-T boundary (Koike, 1996; Sano & Nakashima, 1998; Fig. 3), which is followed by a prominent negative spike of carbon isotope ratio seen both in carbonate and organic carbon (Musashi et al., 2001, 2006 this volume). A similar negative shift is detected in many P-T boundary sections mainly around the Tethyan domain (e.g., Baud et al., 1989; Holser et al., 1989), including the GSSP (Global Stratotype Section and Point) in Meishan, South China (Yin et al., 2001). The detection of the same signal in mid-oceanic paleoatoll limestones indicates the global nature of the isotopic change in oceanic carbon reservoir at the P-T boundary.

Concerning the G-L boundary, another sharp faunal turnover was recently recognized between the Capitanian (Upper Guadalupian) dark-gray limestone and the overlying Wuchiapinian (Lower Lopingian) light gray dolomitic limestone in the Kamura section in Kyushu and the Akasaka section in central Japan (Ota et al., 2000; Isozaki & Ota, 2001; Ota &
The Capitanian limestone is enriched in large-tested verbeekinid fusulines, such as *Yabeina* and *Lepidolina* (up to 1 cm in diameter), whereas the Wuchiapingian rocks lack large ones and contain only much smaller-tested schubertellids, ozawainellids, and staffellids, such as *Codonofusiella* and *Reichelina* (less than 2 mm). The sharp reduction in fusuline tested size in continuous sections may indicate the appearance of an environmental stress to screen out large taxa across the G-L boundary, as will be mentioned later. A dramatic change in lithology coupled with such a fusuline ‘lilliput syndrome’ was detected at two sections in Japan, currently separated from each other by ca. 500 km. This implies that the environmental change was not restricted locally around a single seamount and its surroundings, but was of regional extent over two independent seamounts. Thus, this suggests that the oceanic environment around the carbonate buildups may have changed synchronously and extensively in the superocean.

In addition, preliminary analyses of the carbonate carbon isotope ratio ($\delta^{13}C_{\text{carb}}$) in the G-L boundary section at Kamura in Kyushu revealed a gradual negative shift of ca. 3 per mil, from $+5$ down to $+2$ per mil, around the fusuline-defined G-L boundary (Ota et al., 2002). All the signals in lithology, fossil content and carbon isotope ratio suggest that a major re-organization of oceanography occurred in a short interval across the G-L boundary in the surface of the superocean.

### Felsic tuff

A fine-grained felsic tuff bed immediately below the G-L boundary horizon was recently described by Isozaki & Ota (2001) in the paleoatoll carbonates in the above-mentioned two sections in Japan (Fig. 5). This light greenish grey colored tuff bed is just 1 cm thick at Akasaka and 0.3 cm thick at Kamura. The tuffs are strongly weathered into soft clay at both sections, but they contain several kinds of euohedral crystals, such as quartz and plagioclase. XRD analysis detected an abundance of mixed layered montmorillinite-illite, probably derived from altered volcanic glass. A peculiar bulk chemical composition, determined by XRF, of the tuff shows a low silica content around 50 wt% and high aluminium and titanium contents around 30 and 15 wt% respectively. This probably suggests a secondary silica-leaching by strong weathering, thus preventing determination of the primary bulk composition of the tuff. Nonetheless, the occurrence of euohedral crystals of quartz and feldspars indicates the rhyo-dacitic rather than basaltic nature of the tuff. The fine-grained nature of the tuff probably indicates an air-borne origin.
Fig. 5. Stratigraphic column of G-L boundary intervals in paleo-atoll limestone at Kamura and Akasaka, Japan (modified from Isozaki & Ota, 2001; Ota & Isozaki, 2006). Abbreviations L, Y, R, and C stand for fusulin genera Lepidolina, Yabeina, Reichelina, and Codonosaccus, respectively. Note the occurrence of thin felsic tuff at the G-L boundary horizon.
Despite its thinness, the occurrence of felsic tuff in Kyushu and in central Japan at the same horizon and nowhere else in the limestones is noteworthy. As these tuff-bearing carbonates in Kamura and Akasaka probably represent two independent atoll complexes on paleo-seamounts, this supports the air-borne origin of the tuff. A felsic volcanic ash may have been deposited over an extensive area of the superocean at the G-L boundary.

Also in the deep-sea sequence, the G-L boundary interval with a felsic tuff bed is preserved in the Late Permian accretionary complex in the Akiyoshi belt (Uchiyama et al., 1986; Isozaki, 1987; Nishimura et al., 1989). In accordance with its earlier timing of accretion compared with the contemporary mid-oceanic cherts in the Jurassic accretionary complex, this deep-sea sequence is less siliceous and more argillaceous than typical pelagic chert, reflecting relative proximity to the continental margin under hemipelagic conditions. On the other hand, number of tuff beds is more than 10, and the total accumulated thickness attains several meters (Nishimura et al., 1989). Their geochemistry and fine-grained nature suggest that they were derived from rhyo-dacitic volcanism via airborne processes (Nishimura, 1971).

The occurrence of the fine-grained felsic tuff beds is restricted to the Follicucullus monacanthus (Capitanian) Zone and F. scholasticus–F. ventricosus Zone adjacent to the G-L boundary (Nishimura et al., 1989), although their chronological assignment needs more refinement.

Discussion

On the basis of the datasets described above, I first attempt to correlate the G-L boundary events in shallow-water and deep-sea of Panthalassa, focusing on biotic extinction, change in oceanic conditions, and felsic volcanism. Then, the geological implications with respect to the cause of the G-L boundary event are discussed.

Extinction: Across the G-L boundary, a major faunal change occurred both in fusulines and radiolarians. These two major microfossil groups in the Permian were both severely hit by the G-L boundary event, although they were regarded to have had contrasting habitats in the ocean. Fusulines represent a group of ancient, shallow-marine, warm-water, benthic foraminifera that are regarded to have thrived within the euphotic zone (less than 50 m deep) just like modern benthic foraminifera. In contrast, based on our knowledge of modern counterparts, radiolarians are marine plankton of which habitable zone ranges in ocean from the surface water down to over a thousand meters depth. The Late Paleozoic radiolarians are regarded to have occupied by and large the same habitat in the ocean. Thus fusulines and radiolarians, both from the mid-oceanic domain, monitor environments of very shallow-water and of shallow to moderately deep-water, respectively.

The Early-Middle Permian fusuline fauna mainly consists of large-, robust-tested taxa, i.e. Schwagerinidae and Verbeekinidae. The late Guadalupian fauna in low-latitude, tropical realms including Tethys, Panthalassa and western North America, consist of predominant large-tested fusulines; e.g., Yabeina and Lepidolina, up to 1 cm in diameter, with minor amounts of small-tested ones (e.g., Sheng, 1963; Kammer et al., 1973; Ishii, 1990; Ross, 1995). These fusuline giants became abruptly extinct at the G-L boundary, not only in mid-oceanic paleoatoll limestones (Isozaki & Ota, 2001; Ota & Isozaki, 2006) but also in continental shelf carbonates (e.g., Sheng, 1963; Stanley & Yang, 1994; Ross, 1995; Wilde, 2003; Yang et al., 2004). In turn, the Lower Wuchiapingian rocks yield only smaller-tested taxa less than 2 mm in diameter, e.g. Codonofusiella and Reichelina (Figs 4 & 5). This size reduction in fusuline tested occurred sharply at the G-L boundary and essentially remained throughout the Lopingian until the total extinction of fusulines at the P-T boundary.

Permian large-tested fusulines evolved to form highly complicated shell structures including kerotheca, whereas the smaller ones were conservative to retain their simple, primitive tested structures. As discussed by Wilde (2003), Yang et al. (2004) and Ota & Isozaki (2006), the sharp screening of the large-shelled fusulines at the end of the Guadalupian may have been related to the dying-out of their symbiotic algae, thus, suggesting a large-scale environmental change in surface waters. According to the classic ecological observations on modern biota, a dominance of small-sized individuals (Pianka’s r-strategist) generally indicates the appearance of a harsh environment. Possible candidates for the cause of stress include temperature change, nutrient deficiency and/or salinity change, but it is still difficult to identify the cause at present. In any case, the sharp faunal turnover of fusulines across the G-L boundary suggests that a seriously hostile condition had appeared in the surface waters of the superocean. A barren interval between the Yabeina–Lepidolina Zone and Codonofusiella–Reichelina Zone, in particular, may represent such a period of high environmental stress. Besides fusulines, major victims among the Middle Permian shallow-water benthos include rugose corals, brachiopods,
Fig. 6. Schematic diagram of the G-L boundary events observed in deep and shallow mid-oceanic sequences in the ancient accretionary complexes of Japan. Note the biotic turnover in radiolarians and fusulines, the redox change in deep-sea sediments, and a clear negative shift in carbon isotope ratio. The transitional interval around the G-L boundary may represent a period of oceanographic re-organization of the superocean with a strong environmental stress for shallow marine benthos and pelagic plankton. The coeval felsic tuff suggests a possible cause-effect relationship between the large scale felsic explosive volcanism and the boundary mass extinction.

Radiolarians also suffered from a faunal turnover around the G-L boundary, although the long-term deposition of radiolarian chert did not change much. The Early–Middle Permian mid-oceanic radiolarian faunas were dominated by the genus *Pseudoalabaillella*, which had kept its long-term lineage from the Late Carboniferous. Around the G-L boundary, this genus declined, whereas the new genus *Follicucullus* appeared (Ishiga, 1990). This faunal turnover is significant because the long-ranged and long-dominant genus *Pseudoalabaillella* declined in abundance and in size for the first time in the Permian. An unfavorable situation may have emerged for the pre-existing radiolarians in mid-Panthalassa at the end of the Guadalupian, and the vacant niche was taken over by the new genus in the Wuchiapingian. This biotic turnover in radiolarians suggests that a considerable change had taken place not only in surface waters, but also in moderately deep-water of mid-Panthalassa. The Paleozoic radiolarians, just like modern ones, probably had a wider geographical distribution than fusulines, i.e., from low-latitudes even to higher latitudes, and therefore their faunal turnover possibly reflects a major oceanographic reorganization on global scale. The apparent temporal coincidence of the deep-sea redox change provides further evidence for oceanic reorganization at the G-L boundary.

Figure 6 schematically summarizes extinction/radiation patterns of the major microfossils (radiolarians and fusulines), interval of deep-sea anoxia, secular change in carbon isotope ratio, and volcanic event around the G-L boundary on the basis of the deep-sea cherts and shallow-water paleoatoll limestones from ancient mid-Panthalassa. This shows the approximately synchronous faunal turnover both in fusulines and radiolarians and the onset of a redox change in deep-sea waters of the superocean around the G-L boundary, although more work is needed to enhance the resolution in correlating fusuline, radiolaria, and conodont zones to constrain the extinction timing. In a large picture, it seems likely that the oceanic conditions changed considerably from bottom to top of the superocean at the end of the Guadalupian. A unique transitional interval characterized by a strong environmental stress probably have appeared then to terminate pre-existing taxa and to suppress origination/radiation of new comers.

Carbon behavior: A clear negative shift in $\delta^{13}C_{\text{carb}}$ across the G-L boundary was preliminarily detected in the paleoatoll carbonates at Kamura in Kyushu (Ota et al., 2002), and more detailed study is currently 

Y. Isozaki

119
under way. The shift occurs in the uppermost part of the Guadalupian dark gray limestone, immediately above the final occurrence (LAD) of the Guadalupian large-tested fusulines and below the first predominant occurrence (FAD) of the Lopingian dwarfs (Fig. 6). After the high value interval in the Capitanian (around +5 per mil), the $\delta^{13}C_{\text{carb}}$ values start to decrease gradually upsection to +2 per mil in the Lower Wuchiapingian across the G-L boundary. This pattern suggests that the mid-oceanic surface waters changed isotopic composition along with the above-mentioned major reorganization of superocean after the extinction of large fusulines. The interval of the negative carbon isotope shift corresponds to the above-mentioned transitional interval from the Guadalupian to the Lopingian regime under a high environmental stress.

This chemostratigraphic signal is useful for correlation with other G-L boundary reference sections in the world. Judging from the previous results by Baud et al. (1989), Grossman (1994) and Zakharov et al. (2000), the Guadalupian interval of shelf carbonate is characterized by high $\delta^{13}C_{\text{carb}}$ values of around $+4$ to $+6$ per mil, and the Late Lopingian (Changhsingian) by $+3$ per mil. Recently, Wang et al. (2004) reported results of detailed chemostratigraphic analysis in the GSSP candidate section for the G-L boundary in Laibin, South China. There are some misfits between the two sections they analyzed, but the general trend of 2 to 4 per mil drop in carbon isotope ratio across the G-L boundary is concordant with the result from the paleoatoll limestone in Japan. Regardless of fine-scale fluctuations within both the Guadalupian and Lopingian, the mid-oceanic limestones are correlated with shelf carbonates in South China not only biostratigraphically but also chemostratigraphically. Thus a major reorganization is likely to have occurred in the carbon reservoir in ocean in a global scale, probably associated with the G-L boundary biotic turnover of benthos and plankton.

_Felsic volcanism:_ The occurrence of air-borne rhyodacitic tuff beds both in shallow marine and deep-sea sequences (Fig. 6) indicates that the volcanic ash was deposited over extensive areas of western Panthalassa. The plate tectonic setting of these mid-oceanic rocks requires, for the primary depositional site, ca. 3,000-km separation from Permian Japan, i.e. the eastern continental margin of South China (Fig. 2). Source of the volcanism is unknown at present but the felsic volcanic ash was deposited in the mid-Panthalassa domain, at least in its western half, around the G-L boundary time. The total thickness of the tuff is much greater in the accreted sequence in the Upper Permian accretionary complex than that in the Jurassic complex, suggesting the proximity to the source. Judging from such a difference in thickness and from the modern global atmospheric circulation pattern, the relevant ash probably came from somewhere to the west of Panthalassa.

Isozaki & Ota (2001) and Isozaki et al. (2004) pointed out the good correlation between the above-mentioned mid-oceanic tuff beds and the one at the G-L boundary horizon in South China. The G-L boundary tuff in South China that has been traditionally called the Wangpo bed (Lu, 1956) has long been regarded as a claystone. However, this unit contains abundant volcanogenic material including euhedral felsic phenocrysts (Isozaki et al., 2004). This G-L boundary tuff bed, ca. 1–2 m thick, occurs throughout western South China from Shaanxi to Guanxi, covering nearly 1,000 km in an N-S direction (Li et al., 1991; Isozaki & Ota, 2001; Isozaki et al., 2004). Such a regional coverage within South China requires that the source rhyodacitic volcanism was of considerable magnitude. Moreover, the above-mentioned occurrence of a coeval tuff in mid-oceanic sequences proves much more extensive delivery of the ash. Western Panthalassa, together with South China as a whole, was extensively covered by air-borne tuff of rhyodacitic composition at the G-L boundary. An unusually large-scale and highly explosive volcanism may explain such a long-distance delivery and a huge volume of tuffaceous material.

There is no reported positive evidence ever reported for an extraterrestrial impact at the G-L boundary. On the other hand, the Emeishan Traps (continental flood basalts) in western South China and the Panjal Traps in northern India are often suggested as the cause of the end-Permian mass extinction on account of their chronological similarity (e.g., Chung & Jahn, 1995; Chung et al., 1998; Zhou et al., 2002; Ali et al., 2002). Their basaltic geochemistry, however, can explain neither the violent, explosive eruptions required nor the rhyodacitic nature of the G-L boundary tuff. There are some andesitic basalt and rhyolite units only in the middle and upper parts of the Emeishan Traps (e.g., Xu et al., 2001), but they are too small to account for the entire coverage of the above-mentioned G-L boundary tuff. Recent dating of the Emeishan Traps indicates they range from 259 Ma to 251 Ma (Lo et al., 2002; Hou et al., 2002; Zhou et al., 2002). This Late Permian volcanism also appears too late to have been responsible for the G-L boundary (ca. 260 Ma) tuff deposition and extinction. Thus andesitic basalt and rhyolite of the Emeishan Traps are not likely the source volcanism of the felsic tuff. Also there is not a good candidate for large-scale rhyodacitic volcanism.
neither in the mid-ocean that could have affected the biosphere on such a global scale.

The above observations indicate that the magnitude of the putative G-L boundary volcanism was extraordinarily large, and quite distinct from ordinary plate tectonic-driven volcanism along island arcs or mid-oceanic ridges. Arcs and in particular ridges keep forming as long as plate tectonics is in operation on the Earth, thus have nothing to do with rare and catastrophic events like a mass extinction. Isozaki (2000) speculated that the G-L boundary volcanism might have been related to episodic superplume activity in the Earth's mantle. Highly alkaline volcanism that gives rise to kimberlites and carbonatites may have been responsible for the G-L boundary violent eruptions because the responsible magmas typically carry huge amounts of gas (mostly CO₂ and SO₂; e.g., Bailey & Hamilton, 1990; Ray & Ponde, 1999) to trigger extraordinarily violent eruptions, and also because they may be precursors of large mantle plumes prior to main flood basalt eruptions (Bell, 2001). Morgan et al. (2004) proposed a similar interpretation (the ‘Verneshot hypothesis’) to explain the pseudo-impact evidence from some Phanerozoic mass extinction horizons. Some alkaline plutons in SW China of similar age to the G-L boundary (e.g., the Maomaogou and Panzhihua syenite complexes in southern Sichuan; Co-operative Geological Research Group of Japan and China in the Panxi Region, 1986; Lo et al., 2002; Hou et al., 2002) would be good candidates for such plume-related alkaline igneous activity, although their effusive counterparts have not so far been recognized probably owing to surface erosion. Isozaki (2000) further suggested that the end-Permian biospheric perturbation (and mass extinction) might have been triggered by plume-generated intense volcanism of such an alkaline, felsic to intermediate in the precursory stage of the volcanism rather than from the main flood eruption stage. The input of a huge amount of CO₂ to the atmosphere by such plume-related volcanic activity may well explain the post-extinction global warming and sea-level rise in the Late Permian (e.g., Retallack, 1999; Hallam & Wignall, 1999).

All the previous discussions have focused on the apparent chronological coincidence between the G-L boundary mass extinction and continental flood basalt volcanism, e.g., the Emeishan Traps. This type of approach may suggest a possible causal link between extinction and volcanism but it can never prove it without material-based evidence, as demonstrated in the K-T boundary controversy. On the other hand, the G-L boundary tuff from Panthalassa and South China demonstrates the first geological hard evidence for a possible link between the G-L boundary event and large-scale volcanism. Further geochemical and geochronological analyses are inevitable for the G-L boundary tuff and probably for relevant, coeval igneous rocks in South China and elsewhere.

Conclusion

The recent study on the accreted ancient mid-oceanic rocks from the lost Panthalassa superocean reveals new constraints on the G-L boundary event. Across the G-L boundary, (1) faunal turnover occurred synchronously both in fusulines (sessile benthos in the shallow marine euphotic zone) and radiolarians (plankton in the shallow to much deeper water column), (2) the P-T boundary deep-sea anoxia (superaoxia) started in deep Panthalassa, (3) δ¹³C values started to decrease gradually from +5 to +2 per mil in the superocean surface, and (4) airborne volcanic ash of rhyo-dacitic composition was deposited over western Panthalassa in addition to South China, as summarized in Fig. 6.

The changes in redox and carbon isotope ratios, in particular, indicate that the oceanography of the superocean changed drastically across the G-L boundary. Although the correlation between the shallow and deep-water sequences is still tentative, their apparent temporal coincidence with the major biotic turnover in shallow-water benthos and deeper-water plankton suggests that the whole marine biota in the superocean suffered from a major perturbation of a global scale. The newly identified G-L boundary felsic volcanism may have been responsible for all such environmental turmoil at the time, because it apparently demarcated the long-lasting Guadalupian regime in the shallow and deep ocean from the Lopingian one including the recovery of marine organisms.

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