Fig. 4.11. Photomicrographs of the striped limestone of subunit S3. A) Algal-rich sparitic wackestone, B) algal-poor sparitic grainstone.

Fig. 4.12. Photomicrographs of the striped limestone of subunit S4.

Fig. 4.13. Photomicrographs of the striped limestone of subunit S5. Algal-rich peloidal wackestone, partly sparitic.

Fig. 4.14. Photomicrographs of the boundary between the striped limestone and light gray limestone. Black part (lower) contains algae and a lot of peloids, whereas light gray part (upper) consists of grainstone with algae. There is no greenish bed in the KW section.

Fig. 4.15. Photomicrographs of the light gray limestone of subunit W1. A) fusuline-rich grainstone, B) foraminiferal grainstone.

Fig. 4.16. Photomicrographs of the light gray limestone of subunit W2. A) wacke/lime-mudstone, B) wacke/packstone.

Fig. 4.17. Large scale cross-bedding in Unit S of the Akasaka limestone at the MT section. A: An outcrop view of the striped limestone with cross-bedding. B: A sketch of the same outcrop, showing the clear cross-cutting relationships among black/white beds. C: a polished surface of the cross-bedded limestone of a drilled core sample. It is noteworthy that the high-angle cross-bedding indicates a significantly high-energy depositional setting for this interval.

Fig. 4.18. Polished slab of the boundary between the striped limestone and light gray limestone at KW section. Note that bedding plane is clearly truncated by overlying light gray limestone, and no greenish bed occurs here.

A
light gray limestone (Unit W)
Fig. 4.19B
striped limestone (Unit S)
B
light gray
limestone
greenish tuff
Fig. 4.19C
2 cm
striped
limestone

Fig. 4.19. Photographs of the unique features at the transition from Unit S to Unit W immediately below the G-LB in the MT section of the Akasaka limestone. A: Photograph of the contact of Unit S and Unit W. Note an irregular contact. B: Enlarged photograph of Fig. 4.19A. Note that a thin greenish bed occurs between the two units and that the bedding of the topmost Unit S was clearly truncated by the wavy bed of greenish bed. This suggests the development of a hiatus likely caused by subaerial exposure/erosion at the end of the Capitanian. C: Polished surface of the contact with the unique 5 mm-thick greenish bed. The striped limestone is clearly truncated.

Fig. 4.20. Microscopic picture of boundary between the striped limestone and light gray limestone with intercalating greenish bed. Note that striped limestone does not contain any fusulines, but light gray limestone contains small fusulines (*Nanlingella suzukii*). The contact between the greenish bed and underlying/overlying limestone is wavy.

Fig. 4.21. Microscopic picture of greenish bed. Note that this bed consists of extremely fine-grained matrix and very few minerals.

Fig. 4.22. Quantitative analysis of minerals and matrix included in the greenish bed of the Akasaka Limestone. Each bar indicate the range of element contained.

Fig. 4.23. XRD chart for greenish bed in the Akasaka Limestone (MT section). Apatite (green bars), muscovite (blue bars) and calcite (yellow bars) are detected.

Fig. 4.24. XRF mapping for greenish bed in the Akasaka Limestone.

Fig. 4.25. XRF mapping for greenish bed in the Kamura Limestone.

Fig. 4.26. Rare Earth Element pattern of the greenish beds in Akasaka and Kamura immediately before the G-LB. Normalized by chondrite. Normalized values are after Sun and McDonough (1989).

<table>
<tr><th></th><th></th><th>fusuline zones</th><th>conodont zones</th></tr>
<tr><td rowspan="2">Lopingian
G-LB</td><td rowspan="2">Wuchiapingian</td><td rowspan="2">Codonofusiella
-Reichelina
Zone</td><td>Clarkina dukoensis
Zone</td></tr>
<tr><td>Clarkina postbitteri
postbitteri Zone</td></tr>
<tr><td rowspan="3">Guadalupian</td><td rowspan="3">Capitanian</td><td rowspan="3">Yabeina globosa
Zone</td><td>Clarkina postbitteri
hongshuiensis Zone</td></tr>
<tr><td>Jinogondolella
granti Zone</td></tr>
<tr><td>Jinogondolella
xuanhuanensis Zone</td></tr>
</table>

Fig. 4. 27. Correlation chart of fusuline and conodont zones around the G-LB (Jin et al., 2006).

Fig. 4.28. Stratigraphic column of the KW section in the northern part of the Akasaka Limestone, which spans across the G-LB.

Abbreviations for rock types; M: lime-mudstone, W: wackestone, P: packstone, G: grainstone (the same for Figs. 4.29, 4.30, 4.31, 4.34, 4.35).

Note that the extinction of the large-tested fusulines (*Yabeina*) occurred considerably earlier than the G-LB that is marked by the appearance of the Wuchiapingian assemblage, and that a remarkable facies change has occurred in the latest Capitanian shallow marine limestone deposited on the top of ancient seamount.

Fig. 4.29. Stratigraphic column of the OS section in the Ishiyama Limestone, which spans across the G-LB.

Note that a similar changes in fusuline diversity and limestone facies were recognized also in the Ishiyama limestone. Although this section is separated for ca. 10 km from the Akasaka limestone, almost the identical litho- and biostratigraphy of the two limestones suggests the deposition on the same ancient seamount.

Fig. 4.30. Stratigraphic distribution of the late Capitanian to earliest Wuchiapingian bioclasts in the KW section of the Akasaka limestone.

LO: horizon of the last occurrence for taxon.

Note that the black limestone of Unit B is characterized by abundant/diverse bioclasts. In the overlying units, the abundance and diversity of bioclasts declined remarkably; however, calcareous algae were affected insignificantly. More detailed figure around the barren interval is shown in Fig. 4.31.

calcareous algae
bivalve
gastropod
brachiopod
ostracod
crinoid
large fusuline
small fusuline
smaller forraminifer
Wuchiapingian
C.R. Zone
G-LB
Capitanian
barren interval
Yabeina Zone
W1
20
hiatus
S5
S4
S3
S2
15
S1
10
B4
5
0
m
B3
appearance of Late Permian fusuline assemblage
disappearance of gastropod and smaller foraminifer (second extinction)
disappearance of large fusuline and large bivalve (first extinction)
M W P G

Fig. 4.31. Detailed stratigraphic column of the G-LB interval at the KW section showing the stratigraphic distribution of bioclasts.

Note that shallow marine animals disappeared in two steps; i.e. first at the base of Subunit B4 (base of the barren interval) and second in the lower Subunit S1. The first episode involved large-tested fusulines and large bivalves, whereas the second one terminated gastropod and smaller foraminifers. The upper half of the barren interval is represented by merely minor brachiopods, ostracods and crinoids, except for calcareous algae. The studied interval at 3 sections in the Akasaka-Ishiyama area is divided into 4 distinct periods in terms of environmental stress; i.e., 1) a period of relatively high animal biodiversity (Subunits B1-B3), 2) a period with environmental stress(es) (Subunit B4 to lower Subunit S1), 3) a period under the severest stress(es) with the minimum biodiversity (upper Subunit S1 to lower Subunit W1), and 4) a period of recovering with lesser stress(es) (upper Subunit W1 and above).

Fig. 4.32. Simplified model of seamount and atoll limestone. Note that seamount subsides with time goes on, generally does not uplift.

Fig. 4.33. A: A simplified model of seamount, B: depositional setting around atoll complex. In general, ocean surface is relatively high-energy environment, but in contrast, subtidal setting in lagoon is only a relatively low-energy environment in the mid-superocean.

Fig. 4.34. Secular change in depositional facies of the atoll carbonates during the latest Capitanian and the earliest Wuchiapingian recorded in the Akasaka limestone.

Note that wackestone or packstone was dominant in Subunit B3 and B4, whereas grainstone is dominant in Unit S and Subunit W1. Cross-bedding occurs in Subunit S1, S3 and S5. This overall trend indicates relatively low-energy environment in Subunit B3 and B4 and relatively high-energy environment in Unit S and Subunit W1.

Fig. 4.35. Secular change in depositional facies of the atoll carbonates from the Capitanian to the Wuchiapingian recorded in the Akasaka limestone.

Note that Unit B and Unit W were deposited mostly in the subtidal setting, whereas Unit S and the lowermost Unit W were in the much shallower intertidal setting. This overall trend in depositional setting consists of the pair of shallowing during the latest Capitanian and the deepening during the earliest Wuchiapingian. It is noteworthy that the G-LB corresponds to the timing of the lowest sea-level with surface erosion in the entire atoll history. This pattern is concordant with the reconstructed sea-level curve based on the sequence stratigraphy analysis of the contemporary shelf sediments around the Pangea (Haq and Schutter, 2008).

Fig. 4.36. Ternary plot of La-Th-Sc (after Taylor and McLennan (1985)) for the greenish bed in Akasaka and Kamura. Note that greenish bed both in the Akasaka and Kamura was driven by felsic source, rather than mafic source.

5. Chemostratigraphy

As to the paleoenvironmental studies, analyses of isotopic composition of ancient seawater recorded in sedimentary rocks are inevitable. Carbon, oxygen, and strontium isotope ratios are representative chemostratigraphical proxies frequently used. In general, stable carbon isotope ratio ($\delta^{13}C_{carb}$) is used to monitor the change in primary productivity in past oceans, whereas strontium isotope ratio ($^{87}Sr/^{86}Sr$) is measured for checking continental material flux into oceans reflecting erosion of continental crust.

The Permian period recorded a unique history of seawater C-isotope signature (e.g., Saltzman and Thomas, 2012). Particularly in the Capitanian, an interval of extremely high $\delta^{13}C_{carb}$ values were detected and named the Kamura event on the basis of the detailed analyses on the plaeo-atoll limestone in Japan (Isozaki et al., 2007a, b); however, complete stratigraphic continuity was not guaranteed for the entire Capitanian interval.

Also concerning $^{87}Sr/^{86}Sr$ of seawater, the Permian period recorded a unique history (McArthur et al., 2012). Particularly during the Capitanian, an interval with extremely low $^{87}Sr/^{86}Sr$ values, the lowest of the Phanerozoic, was recognized by previous studies, and it is called "Permian minimum" (e.g., Korte et al., 2006; Kani et al., 2008, 2013). The precise duration of this minimum interval, however, has not been determined.

In order to clarify the background environmental changes in mid-ocean during the critical interval around the G-LB, this study analyzed $\delta^{13}C_{carb}$ and $^{87}Sr/^{86}Sr$ isotope stratigraphy of the Middle-Upper Permian Akasaka limestone, in particular, using the unaltered drilled core samples obtained immediately next to the studied section (see Chapter 3, Fig. 3.3). High-resolution analyses were made for the $\delta^{13}C_{carb}$ and $^{87}Sr/^{86}Sr$ stratigraphy of the Upper Formation of the Akasaka Limestone and the overlying Ichihashi Formation.

5.1. Carbon isotope measurements

5.1.1. Samples and analytical methods

I collected specimens of the Capitanian Upper Mb and Wuchiapingian Ichihashi Fm at KW section. As to the boundary between Unit S and Unit W, extremely high-resolution measurements were done in the MT section. Rocks of the two sections are unmetamorphosed and mostly fresh, and those with strong weathering and with

many calcite veins were screened out in the field and in the laboratory under the microscopic observation.

The micritic part from each horizon was milled by microdrill after examining under the microscope. Approximately 100 mg of the aliquot samples were reacted with 100% H_3PO_4 at 90 °C in an automated carbonate device (Multiprep) coupled with a Micromass Optima mass spectrometer at the Geological Survey of Japan, AIST. Here, $\delta^{13}C = [((^{13}C/^{12}C_{\ sample})/(^{13}C/^{12}C_{\ standard})) - 1] \times 1000$, and $\delta^{18}O = [((^{18}O/^{16}O_{\ sample})/(^{18}O/^{16}O_{\ standard})) - 1] \times 1000$. All isotopic data are reported as per mil (‰) relative to Vienna Pee Dee belemnite (V-PDB) standard. The internal precision was 0.03‰ and 0.04‰ (1σ) for $\delta^{13}C$ and $\delta^{18}O$, respectively, based on replicate measurements of 23 consecutive samples of the NBS-19 calcite standard (Suzuki et al., 2000).

5.1.2. Results

Tables 5.1 and 5.2 list all the measurements of $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ of 147 samples from the 142 m-thick study section. Fig. 5.1 and Fig. 5.2 show a secular change in $\delta^{13}C_{carb}$ values plotted on the stratigraphic column. Especially, Tables 5.2 and 5.3 shows high-resolution measurements for boundary between the striped limestone and light gray limestone in KW and MT section, and Fig. 5.3 shows pictures of each sample.

All $\delta^{13}C_{carb}$ values showed a wide range from -0.377 to + 6.591 ‰. On the other hand, $\delta^{18}O$ values fluctuated between - 2.001 and - 14.867 ‰, which might be partly due to a slight diagenetic alteration. However, based upon the analysis of the correlation between $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$, both data were independent in the secular change, showing certain significant trends (Fig. 5.4). Thus I consider that the $\delta^{13}C_{carb}$ values were not likely affected by secondary alteration but reflect the isotopic composition of the inorganic carbon reservoir in ancient seawater, in which the limestones were deposited.

In the *Yabeina* Zone, $\delta^{13}C_{carb}$ values range from -0.377 to +6.591 ‰. Within the lower half of this zone, $\delta^{13}C_{carb}$ values were mostly less than +4 ‰ except for the 4 samples from the bottom. On the other hand, within the upper half, $\delta^{13}C_{carb}$ values were relatively high more than +4 ‰ and the highest value was up to + 6.5 ‰.

In the barren interval, $\delta^{13}C_{carb}$ values range from +1.986 to +5.878 ‰. The highest

value was placed 1.8 m below the G-LB.

In the *Codonofusiella-Reichelina* Zone, $\delta^{13}C_{carb}$ values were relatively low, ranging from +1.465 to +2.65‰.

As to the G-LB (Fig. 5.3), 10 and 8 measurements were done in the MT and KW section. In the MT section, 7 measurements were done for the striped limestone (Capitanian) and 3 for the light gray limestone (Wuchiapingian) and In the KW section, 5 for the striped limestone (Capitanian) and 3 for the light gray limestone (Wuchiapingian). In the striped limestone, $\delta^{13}C_{carb}$ values range from +2.394 ‰ to +4.086 ‰, whereas in the light gray limestone from +2.707 to +3.470 ‰, slightly lower than those of the striped limestone.

5.2. Strontium isotope measurements

5.2.1. Samples and analytical methods

In order to obtain the primary Sr isotope ratio, fresh samples without secondary diagenesis are needed. Brachiopod shell or conodont, which consist of low-Mg calcite, are preferred for the Sr isotope analysis, but fine-grained micritic limestone also is regarded to have initial Sr isotope ratio of contemporary seawater (Popp et al., 1986; Denison et al., 1994; Denison and Koepnick, 1995; Martin and Macdougall, 1995). In this study, owing to the facies-related scarcity of brachiopods and conodonts in the Permian limestone of Akasaka, we specifically chose fine-grained limestone (micritic part composed of pure carbonates with scarce terrigenous components) for analysis.

I analyzed 29 samples of vein-free fresh samples were analyzed for Sr isotope ratio. I found new outcrop due to the mining at the KW section. I analyzed samples from the new outcrop, whose are lower horizons than the bottom of drilling core. Age was not fully assigned, but probably correlated to Wordian because of the scarceness of *Yabeina*. In addition, especially across the G-LB, high-resolution measurements were done (Fig. 5.6).

For Sr isotope measurements, 40-60 mg of handpicked specimens from each samples were washed in 80 °C Milli-Q water for 3 hours and then dissolved in 3 ml of suprapure 1 M acetic acid. Sr was extracted in 1 ml micro-columns filled with 100 µl Sr Spec resin (ElChrom Industries). The column was rinsed with 3 ml of 3 M HNO_3, and Sr was eluted with 3 ml of H_2O. Separated Sr was loaded on single W filaments with Ta activator. Samples were analyzed by thermal ionization mass spectrometer (TIMS;

Finnigan MAT 262) at the Faculty of Science, Kumamoto University. All data were corrected for internal mass bias using $^{87}Sr/^{86}Sr$ = 0.1194. Recent average value of standard NIST SRM 987 are $^{87}Sr/^{86}Sr$ = 0.710277 ± 25 (2SD; n = 19). Our laboratory blanks were <500 pg.

For the Akasaka Limestone, Sr content is 327 ppm for the black limestone, 450 ppm for the striped limestone, and 53 ppm for the light gray limestone. Mn/Sr ratios are less than 1.0 and Rb/Sr ratios are less than 0.001 for all samples. Therefore, effects of secondary diagenesis for Sr isotope ratio are negligible, and the very low Rb/Sr ratios make the measured $^{87}Sr/^{86}Sr$ ratios of carbonates close to the initial ratios of Permian with negligible age effect.

5.2.2. Results

Tables 5.4 and 5.5 show the measurements of $^{87}Sr/^{86}Sr$ ratio of the carbonate samples collected from Akasaka. Fig. 5.5 displays their stratigraphic change. The $^{87}Sr/^{86}Sr$ values range from 0.7068 to 0.7073. The residence time of Sr in seawater is estimated ca. 3 m.y. in modern oceans. When I discuss past records on the resolution of several tens of million years, therefore, all the ancient seawaters were regarded to have been well mixed isotopically in general.

In the *Neoschwagerina* Zone, $^{87}Sr/^{86}Sr$ ratio range from 0.706996 to 0.707263. The lowermost (M601) were relatively high value, but sharply declined below 0.7070, and then kept slightly above 0.7070.

In the *Yabeina* Zone, $^{87}Sr/^{86}Sr$ ratio range from 0.706853 to 0.707066. Except for one horizon (B30.60), $^{87}Sr/^{86}Sr$ ratio was below 0.7070.

In the barren interval, $^{87}Sr/^{86}Sr$ ratio range from 0.706816 to 0.707072. $^{87}Sr/^{86}Sr$ ratio was extremely low with the lowest value in this section (A33.65).

In the *Codonofusiella-Reichelina* Zone, $^{87}Sr/^{86}Sr$ ratio sharply increased, ranging from 0.707205 to 0.707372.

As to the G-LB (Fig. 5.6), $^{87}Sr/^{86}Sr$ ratio was 0.707003 and 0.706982 12 cm and 2 cm below the boundary. On the other hand, $^{87}Sr/^{86}Sr$ ratio was 0.707237 2 cm above the boundary.

Ancient marine carbonates were often altered secondarily by syn-depositional and post-depositional diagenetic processes, and the primary Sr isotopes values were modified either (Gröcke et al., 2007). I cannot totally neglect the possibility of

secondary diagenetic overprint to the measured samples from the Permian accreted limestones in Japan, as they often suffered from subduction-related multiple deformations. Nonetheless, as diagenetic alteration by meteoric water generally brings Sr isotope values to the higher side. As long as the measured values from the Akasaka limestone are relatively low with respect to the reported values of the Phanerozoic carbonates (Figs. 1.3, 1.4), they likely represent the primary signatures. Even in some interval with relatively scattered data, the lower envelop of the measured values likely approximates the primary signal.

5.3. Discussion

5.3.1. Correlation with the Kamura Limestone

5.3.1.1.Stable carbon isotope profile

According to the newly obtained data from the Akasaka Limestone (Tables 5.1-5.3; Fig. 5.1), the upper Capitanian interval is characterized by extremely high $\delta^{13}C_{carb}$ over +5 ‰, whereas the lower Capitanian interval by slightly lower $\delta^{13}C_{carb}$ values around +4 ‰. In contrast, the Wuchiapingian interval is characterized by clearly lower $\delta^{13}C_{carb}$ values less than +3 ‰. These values indicate that the Permian mid-oceanic seawater around the Akasaka paleo-seamount was enriched in ^{13}C.

Stable carbon isotope ratios were analyzed also in another paleo-atoll limestone, Kamura Limestone in Kyushu, Southwest Japan (Isozaki et al., 2007a, b). Both in Akasaka and Kamura, extremely high $\delta^{13}C_{carb}$ value over +5 ‰ was detected immediately below the G-LB, and then dropped to <+4 ‰ (Fig. 5.7). As the two isolated paleo-atoll complexes originated in mid-Panthalassa shared the same C-isotopic profile, therefore, the documented $\delta^{13}C_{carb}$ records represent an average of low-latitude surface seawater in the mid-superocean. Nonetheless, the stratigraphic extent of the Kamura event, in particular, its onset timing was not determined owing to the imperfect stratigraphic continuity. This study, however, for the first time confirmed that the Kamura event started around the Wordian-Capitanian boundary (Fig. 5.1) and continued throughout the Capitanian. In other words, the Kamura event lasted for nearly 5 million years throughout the entire Capitanian.

5.3.1.2.Sr isotope profile

According to the newly obtained data from the Akasaka Limestone (Tables 5.4, 5.5; Figs. 5.5, 5.8), the Capitanian interval is characterized by extremely low $^{87}Sr/^{86}Sr$ ratio below 0.7070, whereas the Wordian and Wuchiapigian intervals by relatively high values over 0.7070. These values indicate that the Permian mid-oceanic seawater around the Akasaka paleo-seamount was depleted in in ^{87}Sr derived from cntinental sources, as well as the previously reported results from the Kamura paleo-seamount in Kyushu (Kani et al., 2008, 2013). Therefore, the mid-Panthalassan surface seawater shared the same Sr-isotopic composition on a regional scale.

5.3.2. Global correlation

Previous studies have preliminarily documented $\delta^{13}C_{carb}$ records for continental shelf carbonates around Pangea (e.g. west Texas, South China, and Croatia; e.g., Given and Lohman, 1985; Wang et al., 2004; Chen et al., 2011; Saitoh et al., 2013; Isozaki et al., 2011). High $\delta^{13}C_{carb}$ values over +5 ‰ were sporadically reported from the Capitanian limestones in the Delaware basin in West Texas and high-latitude Arctic domains (e.g., Given and Lohman, 1985; Korte et al., 2005), East Greenland (Scholle et al., 1991), and West Spitsbergen (Gruszczynski et al., 1989; Mii et al., 1997). These signals were regarded, however, as local phenomenon that occurred only within restricted basin settings because the values are different from relatively low ones from Tethyan domains (Korte et al., 2005).

Nonetheless, the Kamura event was recently recognized for the first time from European Tethys (Isozaki et al., 2011), and this positively suggests that this event was not of a local context but of global nature. In particular, the paleo-atoll carbonates were deposited in the middle of the superocean, without any physiographical constraints from local basin geometry or from local flux of continental run-off, the isotope data likely reflect a general signature of surface water of the mid-superocean, whereas sediments deposited along continental margins adjacent to lands are usually prone to variable local factors. Thus the records form mid-oceanic paleo-atoll carbonates suggest the averaged seawater isotopic composition of the Permian ocean. Brand et al. (2009) pointed out that the C-isotope records from epeiric seas tend to have a large deviation from those of contemporary mid-oceanic and open shelf environments.

Accordingly, the secular change in $\delta^{13}C_{carb}$ values of the Akasaka Limestone may serve as a global average for mid-superoceanic $\delta^{13}C_{carb}$.

At the GSSP of the G-LB, Penglaitan section, the highest value of over +5 ‰ was detected just at the G-LB, and afterwards, $\delta^{13}C_{carb}$ dropped to +2 ‰ (Wang et al., 2004; Chen et al., 2011). Isozaki et al. (2007a) concluded that the very high $\delta^{13}C_{carb}$ value immediately before the FAD of the Wuchiapingian fusulines (the *Codonofusiella-Reichalina* assemblage) in the Kamura Limestone is correlated with the highest horizon of high $\delta^{13}C_{carb}$ value at the G-LB in the Penglaitan section. These suggest that the G-LB at the Akasaka Limestone is likewise placed at the top of the Unit S, i.e., immediately below the hiatus (Fig. 5.7).

As previously pointed out, the Permian period was characterized by extremely low $^{87}Sr/^{86}Sr$ ratio, the lowest value in the Phanerozoic (e.g. Denison et al., 1994; Korte et al., 2006; McArthur et al., 2012; Fig. 1.3). Korte et al. (2006) reconstructed the secular change in $^{87}Sr/^{86}Sr$ for the Permian, using well-preserved brachiopod samples collected from various places in the world. These previous studies already identified similar isotopic signatures suggesting the Capitanian minimum. On the basis of the present results, therefore, I conclude that the Capitanian seawater was characterized ubiquitously by the extremely low $^{87}Sr/^{86}Sr$ ratios, as low as 0.7068 from the beginning to end of the Capitanian. The $^{87}Sr/^{86}Sr$ values then rose to 0.7072 across the G-LB (Figs. 5.5-5.7).

After all, the secular changes in $\delta^{13}C_{carb}$ and $^{87}Sr/^{86}Sr$ ratio of seawater detected in the mid-oceanic paleo-atoll limestones are fairly correlated with those reported from the Pangean margins. In particular, it is noteworthy that the sharp changes of both isotope proxies occurred across the G-LB; however, the major biodiversity decline started much earlier than the G-LB (Chapter 4).

5.3.3. High productivity in the low-latitude mid-superocean

As the $\delta^{13}C_{carb}$ value generally reflects the isotopic ratio of dissolved inorganic carbon (DIC) in ambient seawater, the secular change in $\delta^{13}C_{carb}$ values measured in the study section is considered to reflect that of DIC in ancient open-ocean water in which the Akasaka Limestone was deposited. The paleo-seamount complex was located in a low-latitude, mid-oceanic domain in the Permian, and carbonates were deposited in a shallow (photic zone) marine setting, therefore, the $\delta^{13}C_{carb}$ values represent the isotopic

signature of the Capitanian to Wuchiapingian surface water around the paleoseamount in the equatorial Panthalassa.

Extremely high $\delta^{13}C_{carb}$ values in the surface mid-superocean indicates the high primary productivity couple with the massive burial of organic carbon (Kump and Arthur, 1999; Saltzman, 2005). As oceanic productivity is strongly controlled by nutrient availability, constant supply particularly of limiting elements, such as phosphorous and nitrogen, is necessary to maintain long-lasting high productivity interval. High riverine run-off, intense upwelling of deep-sea water, or the combination of the two, is likely needed to maintain major high productivity events. In general, low-latitude mid-superocean is situated in oligotrophic environments due to the lack of nutrient supply. A possible mechanism for increasing nutrient supply to the mid-superocean is the active oceanic circulation enhanced by cooling of surface water, as discussed by Isozaki et al. (2007b).

In the Akasaka paleo-atoll limestone, the depositional setting changed into relatively shallower facies before the G-LB, indicating global cooling throughout the Capitanian. Thus this trend can be explained concordantly by the claimed global cooling in the Capitanian.

In contrast, the $\delta^{13}C_{carb}$ values sharply dropped across the G-LB, and the lower Wuchiapingian interval is characterized by relatively low positive values. This likely suggests the collapse of high-productivity probably by the slowing of oceanic circulation under warmer climate. Again this tend is concordant with the reconstructed secular change in sea-level on the basis of sequence stratigraphic analyses (e.g., Haq and Schutter, 2008).

5.3.4. Possible causes of extremely low $^{87}Sr/^{86}Sr$

$^{87}Sr/^{86}Sr$ ratio of seawater is mainly constrained by two sources; 1) Continental riverine flux and 2) hydrothermal flux. The riverine flux has a high $^{87}Sr/^{86}Sr$ ratio ($^{87}Sr/^{86}Sr$ = 0.711), whereas hydrothermal flux has a low $^{87}Sr/^{86}Sr$ ratio ($^{87}Sr/^{86}Sr$ = 0.7035) (Richter et al., 1992). In general, hydrothermal flux does not significantly change through time, thus the continental riverine flux is more effective in the change in $^{87}Sr/^{86}Sr$ ratio of seawater. Thus the extremely low $^{87}Sr/^{86}Sr$ ratios of mid-oceanic seawater in the Capitanian indicate the suppression of continental influx to the superocean for more nearly 5 million years.

The overall secular change in seawater $^{87}Sr/^{86}Sr$ ratio is dependent on the continental configuration, in other words, the pattern of assembly and breakup of supercontinents. During the Paleozoic, all continents moved to form the supercontinent Pangea that resulted in the shortening of total length of coastlines (Korte et al., 2006). According to the widely used paleogeography maps by Scotese (2008), Pangea already finished its assembly by the end of the Carboniferous (ca. 300 Ma). Therefore, the assembly of continents itself cannot be the cause of the Permian minimum in $^{87}Sr/^{86}Sr$ ratio that culminated in the Capitanian.

Another possible explanation is sea-level change. In general, under higher sea-levels, continents tend to be less exposed above the sea-level, thus continental erosion rate is relatively suppressed to minimize the Sr flux into oceans. In the case of the Permian world, the sea-level was kept relatively low owing to the development of the Gondwana glaciation during the Early Permian and of the Kamura cooling event during the Capitanian. In particular, across the G-LB, the lowest sea-level was achieved in the Phanerozoic according to the sequence stratigrapical analyses of continental shelf sediments (Haq and Schutter, 2008), which was clearly confirmed also in the mid-oceanic atoll complex of the Akasaka Limestone in the present study. Therefore, the sea-level change cannot explain the lowest $^{87}Sr/^{86}Sr$ ratio of seawater in the Capitanian.

The fact that global sea-level became the lowest around the G-LB is very important in explaining the lowest $^{87}Sr/^{86}Sr$ ratio of seawater. Global sea-level lowstand means that more seawater is isolated from the ocean and existed on the continent as continental glacier. The expansion of continental glacier is generally induced by global cooling. An interval from the Carboniferous to Early Permian is characterized by Gondwanan glaciation, and this glaciation was previously regarded to have not persisted in the Middle Permian. Recently however, Capitanian glaciers were recognized (Fielding et al., 2008) and this suggests that glaciation might have persisted in the Capitanian. According to these, Kani et al. (2013) speculated that the lowest $^{87}Sr/^{86}Sr$ ratio of seawater in the Capitanian was probably drawn by increased coverage of continental glacier, resulting in suppressed continental erosion rate. After the putative Capitanian glaciation was cancelled, continental erosion rate increased rapidly in the Early Wuchiapingian, resulting in a large flux into the ocean to increase $^{87}Sr/^{86}Sr$ ratio of seawater. In short, the proposed global cooling in the Capitanian and the following

warming across the G-LB can best explain the extremely low $^{87}Sr/^{86}Sr$ ratio of seawater in the Capitanian and the rise afterwards.

5.3.5. Sea-Level change and isotope stratigraphy

As discussed in Chapter 4, the G-LB horizon is likely missing in the Akasaka limestone due to the development of a remarkable hiatus (Fig. 4.19). The pattern in Sr isotope stratigraphy clearly reflected this major environmental change. As shown in Fig. 5.8, the topmost part of the Unit S is characterized by the extremely low Sr-ratios around 0.7068, whereas the lowermost Unit W by relatively high ratios over 0.7071. The clear jump in Sr-ratio for more than 0.003 across the unit boundary is significantly large, thus clearly suggests the existence of a missing stratigraphic interval between the 2 units, which is concordant with the erosional contact at the hiatus.

On the other hand, the secular change in C-isotope apparently looks smooth; nonetheless, there is a small offset in $\delta^{13}C_{carb}$ values (Fig. 5.2). The topmost Unit S has values around +4 ‰, whereas the lowermost Unit W +3.5 ‰or lower. Judging from the erosional contact with a clear gap in Sr-isotope ratio at the hiatus horizon, it is reasonable to expect the discontinuity also in C-isotope profile. As the upper part of Unit S is characterized mostly by high positive values in $\delta^{13}C_{carb}$ around +5-6 ‰, whereas the topmost 10 cm alone has values less than +4 ‰. Within the hiatus interval some unrecognized fluctuations potentially exist; therefore, more detailed analyses is needed for this interval.

Table 5.1. Analytical results of $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ (normalized to the Vienna Pee Dee belemnite) plus TOC in the Akasaka Limestone.

Sample name	**Horizon (m)**	**$\delta^{13}C_{carb}$ (‰VPDB)**	**$\delta^{18}O$ (‰VPDB)**	**TOC (wt %)**
Yabeina Zone				
B69.50	0.4	**+5.127**	**-4.300**	
B67.80	1.8	**+4.331**	**-5.237**	
B64.90	4.1	**+4.655**	**-5.832**	
B64.00	4.8	**+4.058**	**-9.808**	**0.12**
B61.40	6.9	**+2.590**	**-9.890**	
B58.80	9.0	**+3.008**	**-10.314**	
B57.50	10.0	**+2.334**	**-13.410**	
B53.80	13.0	**+3.650**	**-5.107**	
B53.60	13.1	**+3.268**	**-6.059**	
B49.65	16.3	**+2.225**	**-4.539**	**1.10**
B43.90	20.9	**+3.710**	**-4.461**	
B43.75	21.0	**+1.167**	**-4.885**	
B38.10	25.5	**+4.166**	**-6.928**	
B37.20	26.2	**+3.926**	**-5.577**	**0.026**
B34.20	28.6	**+2.606**	**-11.261**	
B33.50	29.2	**+3.672**	**-5.404**	
B28.80	33.0	**-0.377**	**-11.950**	
06AKA2	33.1	**+3.533**	**-6.020**	
B26.80	34.6	**+3.649**	**-4.803**	
06AKA6	38.2	**+4.055**	**-11.275**	
B22.00	38.4	**+3.710**	**-7.276**	
B18.70	41.0	**+3.149**	**-13.394**	
06AKA10	42.6	**+2.233**	**-6.169**	
B16.00	43.2	**+1.153**	**-2.333**	
B13.50	45.2	**+0.905**	**-6.486**	
06AKA14	46.6	**+2.954**	**-7.185**	
06AKA16	47.2	**+3.900**	**-4.743**	

Table 5.1. (continued)

Sample name	Horizon (m)	$\delta^{13}C_{carb}$ (‰VPDB)	$\delta^{18}O$ (‰VPDB)	TOC (wt %)
06AKA17	47.7	**+1.683**	**-4.635**	
B10.15	47.9	**+2.689**	**-5.308**	
06AKA18	48.4	**+4.343**	**-3.993**	
06AKA19	48.8	**+4.215**	**-4.282**	
06AKA21	49.9	**+5.616**	**-4.887**	
06AKA22	50.6	**+5.557**	**-9.221**	
B6.55	50.8	**+3.408**	**-7.524**	
06AKA23	51.0	**+5.852**	**-4.288**	
06AKA24	51.3	**+4.498**	**-11.074**	
06AKA26	52.9	**+4.036**	**-5.827**	
B2.40	54.1	**+4.689**	**-4.109**	
06AKA27	54.2	**+4.247**	**-5.218**	
06AKA30	56.0	**+6.591**	**-3.862**	
06AKA32	57.9	**+2.052**	**-4.157**	
06AKA34	60.5	**+2.519**	**-14.867**	**0.050**
06AKA36	63.4	**+3.711**	**-13.063**	
06AKA37	65.0	**+2.981**	**-7.509**	**2.35**
06AKA38	66.0	**+5.438**	**-4.125**	
06AKA39	66.7	**+4.254**	**-5.616**	
06AKA40	67.6	**+3.749**	**-5.545**	**0.066**
06AKA41	68.3	**+4.233**	**-7.627**	
06AKA42	69.2	**+4.562**	**-5.031**	
06AKA43	70.0	**+3.670**	**-6.311**	
06AKA44	70.9	**+5.416**	**-5.227**	
06AKA45	72.2	**+4.456**	**-10.313**	
06AKA46	72.9	**+4.510**	**-7.298**	
06AKA47	73.7	**+2.209**	**-8.277**	
07AKA201	74.6	**+5.538**	**-3.734**	**0.44**
06AKA48	75.2	**+2.727**	**-8.162**	

Table 5.1. (continued)

Sample name	Horizon (m)	$\delta^{13}C_{carb}$ (‰VPDB)	$\delta^{18}O$ (‰VPDB)	TOC (wt %)
07AKA202	75.6	**+5.336**	**-4.929**	
06AKA49	76.1	**+4.552**	**-8.202**	
07AKA203	76.7	**+4.514**	**-9.112**	
06AKA50	77.2	**+3.044**	**-10.290**	
07AKA204	78.4	**+5.186**	**-5.788**	
A82.65	78.6	**+1.657**	**-11.405**	
07AKA205	79.8	**+4.912**	**-4.822**	
07AKA206	81.0	**+3.856**	**-9.983**	
A78.55	81.9	**+2.811**	**-9.239**	
07AKA207	82.0	**+3.367**	**-8.519**	
A74.40	85.2	**+3.589**	**-10.818**	
07AKA209	86.2	**+4.330**	**-8.264**	
07AKA210	87.0	**+4.530**	**-4.019**	
07AKA213	88.8	**+0.487**	**-11.412**	
07AKA253	88.9	**+2.735**	**-14.735**	
07AKA254	89.6	**+3.830**	**-6.312**	
A67.90	90.4	**+3.118**	**-8.194**	
07AKA256	91.2	**+3.993**	**-4.730**	
07AKA258	92.3	**+3.482**	**-8.884**	
A63.50	93.9	**+4.289**	**-4.979**	
07AKA260	96.7	**+3.822**	**-8.310**	
A59.70	97.0	**+4.328**	**-5.638**	
A56.85	99.2	**+4.248**	**-6.831**	
A54.90	100.8	**+4.804**	**-4.746**	**0.10**
A51.85	101.2	**+5.908**	**-4.660**	
A49.80	102.9	**+4.135**	**-9.205**	
barren interval				
A49.40M	103.2	**+3.825**	**-7.348**	

Table 5.1. (continued)

Sample name	Horizon (m)	$\delta^{13}C_{carb}$ (‰VPDB)	$\delta^{18}O$ (‰VPDB)	TOC (wt %)
A49.40S	103.2	**+5.242**	**-5.038**	
A49.10	103.4	**+2.150**	**-7.180**	
A48.60	103.8	**+4.160**	**-5.569**	
A48.50	103.9	**+4.273**	**-4.032**	
A46.50	105.5	**+4.865**	**-4.010**	
A44.80	106.9	**+4.711**	**-2.890**	
A44.30	107.3	**+4.625**	**-4.049**	
A43.50	107.9	**+4.456**	**-5.469**	
A42.50	108.7	**+4.943**	**-4.541**	
A41.50	109.5	**+4.880**	**-4.090**	
A41.00	109.9	**+4.586**	**-4.926**	
A40.60	110.2	**+4.120**	**-5.283**	
A39.80	110.9	**+3.618**	**-4.183**	
A38.80	111.7	**+3.878**	**-5.358**	
A38.50	111.9	**+3.952**	**-7.558**	
A37.50	112.7	**+4.423**	**-4.080**	
A36.90	113.2	**+4.195**	**-5.595**	
A36.80	113.3	**+2.849**	**-4.552**	
A35.60	114.2	**+3.749**	**-2.193**	
A34.40	115.2	**+4.861**	**-5.411**	
A33.50	115.9	**+3.984**	**-5.356**	**0.0030**
A33.15	116.2	**+5.284**	**-4.764**	
A32.60	116.6	**+4.403**	**-3.341**	
A31.70	117.4	**+1.986**	**-7.640**	
A31.50	117.5	**+4.348**	**-4.152**	
A31.00	117.9	**+3.590**	**-4.035**	
A29.90	118.8	**+4.563**	**-3.070**	
A29.20	119.4	**+5.024**	**-4.416**	
A28.30	120.1	**+5.878**	**-4.971**	

Table 5.1. (continued)

Sample name	Horizon (m)	$\delta^{13}C_{carb}$ (‰VPDB)	$\delta^{18}O$ (‰VPDB)	TOC (wt %)
A28.10	120.2	**+4.666**	**-4.003**	
A28.00	120.3	**+4.723**	**-3.372**	
A27.80	120.5	**+5.492**	**-5.463**	
A27.60B	120.6	**+3.098**	**-4.069**	
A27.60W	120.6	**+1.872**	**-5.080**	
A27.30	120.9	**+3.549**	**-2.809**	
A27.05	121.1	**+3.572**	**-2.897**	
A26.80	121.3	**+3.558**	**-3.901**	
A26.50	121.5	**+3.200**	**-2.987**	
Codonofusiella-Reichelina Zone				
A24.00	123.5	**+2.081**	**-5.238**	
A22.30	124.9	**+2.992**	**-4.682**	
A21.00	125.9	**+2.650**	**-5.015**	
A19.00	127.5	**+1.987**	**-3.778**	**0.00077**
A16.05	129.9	**+1.465**	**-5.959**	
A15.80	130.1	**+1.146**	**-4.935**	
A10.20	134.6	**+2.742**	**-4.008**	
A4.90	138.8	**+2.258**	**-4.735**	**0.00032**

Table 5.2. Analytical results of $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ (normalized to the Vienna Pee Dee belemnite) around the G-LB in the KW section.

Sample name	**Horizon from G-LB (cm)**	**$\delta^{13}C_{carb}$ (‰VPDB)**	**$\delta^{18}O_{carb}$ (‰VPDB)**
striped limestone			
AK-KW-1		**+3.466**	**-2.386**
AK-KW-2		**+3.831**	**-2.355**
AK-KW-3		**+3.777**	**-3.524**
AK-KW-4		**+3.023**	**-2.716**
AK-KW-5		**+3.897**	**-3.471**
light gray limestone			
AK-KW-6		**+2.979**	**-3.116**
AK-KW-7		**+3.470**	**-2.001**
AK-KW-8		**+3.004**	**-3.979**

Table 5.3. Analytical results of $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ (normalized to the Vienna Pee Dee belemnite) around the G-LB in the MT section.

Sample name	**Horizon from G-LB (cm)**	**$\delta^{13}C_{carb}$ (‰VPDB)**	**$\delta^{18}O_{carb}$ (‰VPDB)**
striped limestone			
AK-MB-1	-15.0	**+2.394**	**-6.059**
AK-MB-2	-14.0	**+4.086**	**-4.313**
AK-MB-3	-9.0	**+3.702**	**-5.134**
AK-MB-4	-5.0	**+3.904**	**-4.537**
AK-MB-5	-4.0	**+3.062**	**-5.975**
AK-MB-6	-1.5	**+2.787**	**-5.703**
AK-MB-7	-1.0	**+3.755**	**-5.130**
light gray limestone			
AK-MB-8	+3.0	**+3.154**	**-2.592**
AK-MB-9	+7.0	**+2.707**	**-3.214**
AK-MB-10	+16.0	**+2.793**	**-2.380**

Table 5.4. Analytical results of $^{87}Sr/^{86}Sr$ normalized to the NIST987 of the Akasaka Limestone.

Sample name	Horizon (m)	$^{87}Sr/^{86}Sr$	2SD	Rb/Sr	Mn/Sr
Neoschwagerina Zone					
M601	-31.5	0.707256	0.000027	0.00013	0.00050
M603	-29.2	0.706980	0.000026	0.00004	0.01167
M606	-22.6	0.707022	0.000027	0.00007	0.10470
M614	-13.2	0.707018	0.000026	0.00001	0.00946
M622.5	-5.2	0.707072	0.000029	0.00004	0.00088
Yabeina Zone					
B66.45	2.8	0.706891	0.000027	0.00009	0.00285
B64.35	4.5	0.706976	0.000026		
B49.65	16.3	0.706918	0.000027		
B41.80	22.6	0.706915	0.000025		
B30.60	31.6	0.707065	0.000024		
B16.00	43.2	0.706964	0.000029		
B0.80	55.4	0.706907	0.000029		
A84.25	77.3	0.706938	0.000043		
A75.90	84.0	0.706936	0.000037		
A65.70	92.2	0.706898	0.000027		
A54.70	101.0	0.706837	0.000028	<0.00001	0.00004
A51.45	101.6	0.706958	0.000025	0.00003	0.00037
barren interval					
A49.30	103.3	0.706947	0.000026	0.00006	0.00686
A47.45	104.8	0.707006	0.000025	0.00002	0.00565
A41.65	109.4	0.707055	0.000025	0.00001	0.00375
A37.75	112.5	0.706927	0.000026		
A33.65	115.8	0.706799	0.000026	<0.00001	0.00019
MB09	121.8	0.706923	0.000026	<0.00001	0.00081

Table 5.4. (continued)

Sample name	Horizon (m)	$^{87}Sr/^{86}Sr$	2SD	Rb/Sr	Mn/Sr
Codnofusiella-Reichelina Zone					
A24.30	123.3	0.707197	0.000026	0.00001	0.01189
A22.35	124.8	0.707283	0.000026	0.00005	0.02566
A12.00	133.1	0.707245	0.000026	0.00018	0.01016
A8.40	136.0	0.707189	0.000026	0.00016	0.70829
A4.60B	139.0	0.707387	0.000038		
A4.60W	139.0	0.707370	0.000026	0.00011	0.66524

Table 5.5. Analytical results of $^{87}Sr/^{86}Sr$ normalized to the NIST987 of the Akasaka Limestone.

Sample name	Horizon From G-LB(cm)	$^{87}Sr/^{86}Sr$	2SD
MT-S1	-12	0.707002	0.000033
MT-S2	-2	0.706981	0.000026
MT-W1	+2	0.707236	0.000034
MT-W2	+12	0.707293	0.000032

Fig. 5.1. Chemostratigraphy of stable carbon isotope of carbonates of the KW section in the Akasaka Limestone.

Fig. 5.2. Chemostratigraphy of stable carbon isotope of carbonates around the G-LB in the KW section.

Fig. 5.3. Photograph of G-LB sample of A: MT section, and B: KW section. Each circle shows measured point with $\delta^{13}C_{carb}$ value. Note that the underlying striped limestone (Capitanian) is clearly truncated by overlying light gray limestone (Wuchiapingian).

Fig. 5.4. Crossplot of $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$.

Fig. 5.5. The secular change in $^{87}Sr/^{86}Sr$ ratio of bulk carbonates in the Akasaka Limestone.

Fig. 5.6. Measured point for $^{87}Sr/^{86}Sr$ ratio around the G-LB at the MT section.

Fig. 5.7. Correlation of $\delta^{13}C_{carb}$ between mid-oceanic limestone (this study + Isozaki et al., 2007a,b) and shelf limestone (Wang et al., 2004; Chen et al., 2011).

Fig. 5.8. $^{87}Sr/^{86}Sr$ of the Akasaka Limestone with data of Kani et al. (2013).

6. Synthesis

I analyzed high-resolution litho-, bio-, and chemostratigraphy in the mid-oceanic limestone to clarify the global environmental change around the G-LB and the cause of the G-LB extinction. On the basis of the above results and discussion, I will discuss the cause of the G-LB extinction.

6.1. Global cooling in the late Guadalupian

Our present results for the first time identified the mid-oceanic sea-level drop at the end of the Middle Permian; however, its cause is still unknown. At any option, in order to drive a remarkable regression that affects mid-oceanic realms, a large amount of seawater is definitely removed from oceans and dumped onto lands in the form of continental ice. In other words, a global cooling is essential to cause a large-scale regression of a global context.

$\delta^{13}C_{carb}$ pattern is concordant with the global cooling in the Capitanian and warming afterwards. $^{87}Sr/^{86}Sr$ is also supportive for global cooling in the Capitanian and following warming across the G-LB. The $\delta^{13}C_{carb}$ and $^{87}Sr/^{86}Sr$ patterns in the mid-oceanic limestone, but also in shelf limestones (e.g. ;Isozaki et al., 2011; McArthur et al., 2012), suggesting the cooling in the Capitanian occurred in global context.

6.2. Extinction by cooling

The appearance of global cooling episode in the late Middle Permian was already pointed out by some researchers from the various aspects; e.g. 1) stable carbon isotope stratigraphy (Isozaki et al., 2007a, b, 2011), 2) glacial deposits (Fielding et al., 2008; Fujimoto et al., 2012), and 3) fossil assemblages (Shen and Shi, 2002; Isozaki, 2006, Isozaki and Aljinovic, 2009). In particular, the appearance of a relatively cool climate in the tropical domain, in association with eutrophication, was proposed as a possible cause for the collapse of photosymbiotic systems (Isozaki, 2009; Isozaki and Aljinovic, 2009), and this likely explains the pattern of fossil occurrence across the G-LB in the Akasaka and Ishiyama limestones as demonstrated in this study. In this regard, the putative global cooling (Isozaki et al., 2007, 2011) remains as a promising candidate for the latest Capitanian severest environmental stress. After the G-LB cooling, surviving biota re-dominated under relatively warmer climate might radiate to re-build the

Wuchiapingian community even in remote mid-oceanic atoll settings.

Fig. 6.1. A schematic diagram showing the Late Capitanian to earliest Wuchiapingian environmental changes in the low-latitudes mid-Panthalssan paleo-atoll complexes; the Akasaka and Ishiyama limestones in central Japan. The barren interval likely represents a period under severe environmental stress(es); in particular, its upper half corresponds to that of the worst conditions for the Capitanian shallow-marine protists and animals. The latest Capitanian shallowing of the depositional setting was culminated at the subaerial exposure near the G-LB. This suggests the appearance of a cool climate at the end of Capitanian. The putative cool climate in the Capitanian (the Kamura event; Isozaki et al., 2007) was a likely cause of the extinction of Middle Permian shallow marine fauna in low-latitude mid-Panthalassa.

7. Conclusions

The present litho-, bio- and isotope stratigraphical analyses on the Middle-Upper Permian Akasaka and Ishiyama limestones in central Japan clarified the following new aspects of the extinction-related environmental changes in the low-latitude shallow mid-Panthalassa.

1) Most of the Capitanian to Wuchiapingian shallow-marine limestones were deposited under subtidal settings within the paleo-atoll complex, likely in a lagoon environment, on the top of the Akasaka paleo-seamount.

2) During the terminal Capitanian and the earliest Wuchiapingian around ca. 260 Ma, the sea-level changed significantly. In the late Capitanian, sea-level dropped to expose the top of atoll complex above the sea-level.

3) The abundance of shallow-marine protists (fusulines, smaller foraminifers) and metazoan animals (mollusks, brachiopods, crinoids, ostracods etc.) declined in two steps during the latest Capitanian.

4) The poorest fossil assemblage from the topmost Capitanian, solely composed of ostracods, crinoids, brachiopods, and calcareous algae, likely represents the biota under the strongest environmental stresses in the Akasaka paleo-atoll history.

5) The recovery of fusulines occurred immediately after the onset of the earliest Wuchiapingian sea-level rise.

6) $\delta^{13}C_{carb}$ was high (>+5‰) in the Capitanian, then dropped across the G-LB.

7) $^{87}Sr/^{86}Sr$ was stably low (<0.7070) throughout the Capitanian, and increased to 0.7073 across the G-LB.

These facts recorded a unique episode on the Middle-Upper Permian atoll complex deposited in low-latitude mid-Panthalassa. The Akasaka limestone experienced an overall shallowing during the Capitanian, which culminated at the subaerial exposure/erosion across the G-LB. The development of a clear hiatus at the top of a mid-oceanic seamount indicates that a remarkable sea-level drop has occurred during the latest Capitanian. This is the first confirmation of the G-LB sea-level drop on global scale down to the Phanerozoic minimum, which was suggested by sequence stratigraphy of continental shelf sediments. This suggests that a cool climate likely appeared even in the low-latitude domains in the Capitanian Panthalassa, therefore, the tropical biota were no doubt affected significantly. The Wuchiapingian biota first

appeared immediately after this erosional episode, i.e., during the onset of warming after the G-LB. Isotope patterns of $\delta^{13}C_{carb}$ and $^{87}Sr/^{86}Sr$ are also concordant with cool climate in the Capitanian and warming across the G-LB. All lines of evidence from litho-, bio-, and chemostratigraphic analyses of the present study indicate that the end-Guadalupian extinction occurred in the later half of the Capitanian when a global cooling occurred to lower the sea-level dramatically.

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