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GEOLOGIC BACKGROUND

The fossiliferous Valiabad and Dalir sections, 60 km north of Tehran in the Elburz Mountains of northern Iran, preserve sedimentary records from the late Proterozoic through to the Cambrian (Fig. 1). The Soltanieh Formation, containing the PC-C boundary, consists of carbonates and shales (Hamdi et al., 1989; Brasier et al., 1990; Kimura et al., 1997) (Fig. 1). The Dolomite Member, characterized by peritidal flat carbonate, along with microbial structures of irregular laminae with fenestrae and rhythmic lamination produced by tidal currents, indicates a shallow water and tidal flat environment. The Shale Member is characterized by rare slumps and rhythmically interbedded carbonate layers (<1 m) consisting of wackestone and mudstone with fossil spicules, indicating an open marine subtidal environment. Hamdi et al. (1989) proposed the PC-C boundary at the top of the Lower Shale Member (LSM), where diverse early skeletal fossil assemblages such as *Protohertzina anabarica* appear. On the other hand, "Proterozoic-type" giant acritarchs such as *Chuaria circularis* (Hamdi et al., 1989) disappear in the middle part of the LSM. However, interbasinal stratigraphic correlation cannot rely solely on fossil distributions that are strongly influenced by sedimentary facies. The sequence boundary at the Neoproterozoic-Manykayan (Nemakit-Daldynian) boundary in Siberia (Pelechaty et al., 1996) are correlated to that at the top or in the upper part of the Lower Dolomite Member (LDM) in Iran (Fig. 1), where several shelly fossils such as *Hyolithellus* sp. occur (Hamdi et al., 1989). This correlation does not conflict with $\delta^{13}C$ stratigraphy and biostratigraphy (Brasier et al., 2000). This stratigraphic level can also be correlated to the sequence boundary in member 2 of the Chapel Island Formation at the PC-C stratotype in Newfoundland, approximately 83 m above the PC-C boundary (Myrow and Hiscott, 1993), which also does not conflict with fossil distributions there (Narbonne et al., 1987). The sequence boundary can also be correlated to that at the base of the Uratanna Formation in South Australia (Mount and McDonald, 1992).

The $\delta^{13}C$ values of carbonate and organic matter exhibit covariation throughout the successions (Fig. 1) (Kimura et al., 1997), which indicates a primary isotopic signature (Knoll

et al., 1986). Both of the sections examined here show similar $\delta^{13}C$ curves. A remarkable negative excursion of $\delta^{13}C$ reaches a minimum of -7‰ or -9‰ in the lower half of the LSM, followed by a positive excursion in the organic phase reaching a maximum of -25‰ in the upper half. A negative $\delta^{13}C$ excursion is recognized worldwide in shallow water carbonates around the PC-C boundary (Kimura et al., 1997), e.g., -10‰ in northwest Canada (Narbonne et al., 1994), -4‰ in northeast Siberia (e.g. Pelechaty et al., 1996) and Morocco (Magaritz et al., 1991), -5‰ in Mongolia (Brasier et al., 1996), -8‰ in northern India (Banerjee et al., 1997), -3‰ in Oman (Burns and Matter, 1993; Brasier et al., 2000), and -34‰ in organic carbon in Poland (Strauss et al., 1997)[1]. The global extent of the short-lived negative $\delta^{13}C$ excursion indicates ^{13}C-depletion in shallow marine waters worldwide (e.g., Kimura et al., 1997).

METHOD

 Whole-rock powdered shales were dissolved using HCl-HNO_3-HF-$HClO_4$, then concentrations of Th, U, V, Sc were determined with Yokogawa PMS200 Inductively Coupled Plasma Mass Spectrometer (ICP-MS) at Geological Survey of Japan. Overall uncertainties are typically within ±6%.

 $\delta^{13}C$ values from Garmab in Iran and Yunnan in China were determined on Finnigan Mat Delta E type isogas mass spectrometer at University of Tokyo. Analytical procedures are described in Kimura et al. (1997). Overall uncertainties are typically within ±0.1 ‰.

 Ichnofabric data described in Figure 1 were rigorously analyzed following these schemes and using plastic-impregnated slabs and thin sections of shales. Analysis was operated optically to segregate later diagenetic structures, such as stylolites and alignment of recrystallized clay minerals, from original sedimentary textures. Errors of determination occur in samples having intermediate features between two adjacent indices or very little variation in grain size. Each sample was examined three times independently. If all three results were not the same, these samples, approximately 10% of all, were further carefully examined for final determination in comparison with other samples typical of the indices. Errors of first three examinations did not exist over two adjacent indices; overall uncertainties are therefore

no more than ±0.5 index. For observations at the macroscopic scale, a previous scheme designed for micritic carbonates (Droser and Bottjer, 1986) was not used here, partly because this observation is an order of magnitude smaller than the previous scheme.

EXTENT OF THE OXYGEN DEFICIENT OCEAN

No Th/U data have been reported from the same stratigraphic level in other locations, which does not allow to confirm whether the δ^{13}C-Th/U correlation is a global trend or restricted to the paleo-Tethys. However, reported lithostratigraphic and geochemical data helps to understand U concentration in shales at this stratigraphic level and extent of the oxygen-deficient ocean. Authigenic uranium resides in organic matter, and various Phanerozoic sediments show good positive correlation between uranium and organic contents (Myers and Wignall, 1987; Wignall, 1994; Jarvis et al., 1994). Uranium also shows strong enrichment in phosphorites, and there is a strong positive correlation between U and P contents (Jarvis et al., 1994). Both of organic-rich and phosphatic sediments, therefore, imply enhanced precipitation of uranium from seawater. Around the paleo-Tethyan area, such evidence includes organic rich cherts in the Fara Formation, Oman Mountains (Brasier et al., 2000); the U-Formation, characterized by high content of uranium, in the Ara Group, south Oman (Brasier, written communication); and phosphatic mudstones and siltstones in the middle part of the Uratanna Formation, South Australia (Mount and McDonald, 1992). All of these are stratigraphically correlated to the negative δ^{13}C anomaly at the PC-C boundary. In Poland, widely separated from the paleo-Tethys, organic-rich sediments occur with a negative $\delta^{13}C_{org}$ anomaly around the PC-C (Fig. B after Strauss et al., 1997). Decreased diversity in trace fossils, abundant low-diversity bacterial micro biotas and cyanobacterial microfossils also characterize this stratigraphic level, which was interpreted to reflect dysarerobic/anoxic environments (Strauss et al., 1997). These observations all together indicate global nature to the enhanced uranium precipitation from seawater during the period and oceanic oxygen deficiency. Moreover, in Poland, the negative $\delta^{13}C_{org}$ anomaly coincides with high organic C/P ratios (Fig. B). Strauss et al. (1997) attributed the high C/P to extensive reworking of organic matter with preferencial release of P relative to C through the water column, which has been considered as

typical before the Cambrian explosion (Logan et al., 1995). However, an alternative explanation is enhanced organic phosphorus regeneration that occurs under relatively oxygen-deficient water column (e.g., Ingall and Jahnke, 1997). Further evidence on the organic C/P bleow this stratigraphic level will confirm origin of the high organic C/P.

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ADDITIONAL EXPLANATION FOR Figure.1

Circles—dolomite, squares—calcite, lozenges—mixture of dolomite and calcite, upward triangles—organic carbon in carbonates and downward triangles—organic carbon in shales. Solid symbols—well-preserved original carbonate fabrics, half-filled symbols—moderately altered fabrics, and open symbols—intensely altered fabrics. Curves for $\delta^{13}C$ were drawn using values of both a carbonate and a organic phases (dark curves), or values of either carbonate or organic phase (light curves) ($\delta^{13}C_{carb}$ - $\delta^{13}C_{org}$ = 28.9 ‰).

FIGURE CAPTIONS

Figure A. Plots of V vs. Al, Ti, Nb, Th, and Sc abundances in examined shales in Iran. All elements are insoluble with very short residence time in the ocean (Taylor and McLennan, 1985). R^2 = correlation coefficient.

Whole-rock powdered shales were dissolved using HCl-HNO_3-HF-$HClO_4$, then concentrations of V, Nb, Th, and U were determined with Yokogawa PMS200 Inductively Coupled Plasma Mass Spectrometer (ICP-MS) at Geological Survey of Japan. Overall uncertainties are typically within ±6%. Whole-rock powdered shales were fused and and concentrations of for Al_2O_3, TiO_2 were measured with Rigaku 3270 X-ray fluoresence (XRF) analyzer at Ocean Research Institute, University of Tokyo. Overall uncertainties are typically within ±0.35 and 0.40%, respectively.

Figure B. Variations in $\delta^{13}C_{org}$, TOC and organic C/P ratios around the PC-C boundary at the Bialopole IG-1 drill core in East European Platform, southeast Poland, after Strauss et al. (1997). Three negative (or positive) $\delta^{13}C$ cycles, identified by I, II and III, are visible. Strauss et al. (1997) interpreted the PC-C boundary at 2826m on the basis of the lowest occurrences of a diagnostic early Cambrian acritarch assemblage, 16m below the first occurrence of *Treptichnus pedum*. However, *Vendotaenia*, which disappears around 2826m (Strauss et al., 1997), is not diagnostic to the Neoproterozoic but disappears in the lowest Cambrian at the stratotype in Newfoundland (Narbonne at al., 1987) and in the Manykayan in northern Iran. These observations, together with an apparent sequence boundary in the Bialopole Formation (Strauss et al., 1997) and an underlying tuff bed having yielded a U-Pb SHRIPM age of 551±4.0 Ma (Compston et al., 1995), suggests that the PC-C boundary is slightly below the $\delta^{13}C$ profile and therefore the negative $\delta^{13}C$ anomaly, identified as I, is correlated to that in the LSM, Soltanieh Formation, northern Iran.