The Precambrian-Cambrian Transition in the Southern Great Basin, USA

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ABSTRACT: The Precambrian-Cambrian boundary presents an interesting stratigraphic conundrum: the trace fossil used to mark and correlate the base of the Cambrian, *Treptichnus pedum*, is restricted to siliciclastic facies, whereas biomineralized fossils and chemostratigraphic signals are most commonly obtained from carbonate-dominated sections. Thus, it is difficult to correlate directly between many of the Precambrian-Cambrian boundary sections, and to assess details of the timing of evolutionary events that transpired during this interval of time. Thick sections in the White-Inyo region of eastern California and western Nevada, USA, contain mixed siliciclastic-carbonate lithofacies, and therefore promote correlation between these classic, well-studied lithologic end-members. An integrated stratigraphic approach was applied to the White-Inyo succession, combining lithologic, paleontologic, and chemostratigraphic data, in order to address the temporal framework within the basin, and to facilitate worldwide correlation of the boundary. Results from the southern Great Basin demonstrate that the negative $\delta^{13}C$ excursion that is ubiquitous in carbonate-dominated successions containing small shelly fossils occurs within stratigraphic uncertainty of the first occurrence of *T. pedum*. This global geochemical marker thus provides a link with the primary biostratigraphic indicator for the Precambrian-Cambrian boundary.

INTRODUCTION

The Precambrian-Cambrian (PC-C) transition records one of the most important intervals in the history of life, because it encompasses the appearance and diversification of metazoans, the invasion of the infaunal realm, the advent of biomineralization and predation, as well as dramatic isotopic and atmospheric changes (Lipps and Signor, 1992; Bengtson, 1994; Knoll and Carroll, 1999; Bottjer et al., 2000; Knoll, 2000; Babcock et al., 2001; Fig. 1). Here we draw a distinction between the PC-C *transition*, represented by the post-glacial terminal Proterozoic through the early Cambrian (ca. 600-520 Ma), and the PC-C *boundary*, a chronostratigraphic boundary represented by a point in rock (Global Standard-stratotype Section and Point, or GSSP) in the Fortune Head, Newfoundland, section (Landing, 1994). The GSSP section is composed predominantly of siliciclastics (Narbonne et al., 1987), and the fossil chosen to coincide with the boundary, *Treptichnus* (or *Phycodes*) *pedum*, is restricted to siliciclastic facies. *T. pedum* recently has been demonstrated to occur ~ 4 m below the GSSP (Gehling et al., 2001). Approximately 70% of all PC-C boundary successions are siliciclastic (Landing, 1994). However, many carbonate successions around the world have been more intensely studied because they record the advent of widespread biomineralization and easily obtainable $\delta^{13}C$ chemostratigraphic records (summarized in Kaufman et al., 1997; Shields et al., 1997; Bartley et al., 1998; Shields, 1999; Corsetti and Hagadorn, 2000; Shen and Schidlowski, 2000). Due to endemic biotas and facies control, it is difficult to correlate directly between siliciclastic- and carbonate-dominated successions. This is particularly true for the PC-C boundary interval because lowermost Cambrian biotas are highly endemic and individual, globally distributed guide fossils are lacking (Landing, 1988; Geyer and Shergold, 2000).

Determination of a stratigraphic boundary generates a large amount of interest because it provides scientists with an opportunity to address a variety of related issues, including whether the proposed boundary position marks a major event in Earth history. Sometimes the larger-scale meaning of the particular boundary can be lost during the process of characterization. This is demonstrated in a plot of PC-C boundary papers through time (Fig. 2): an initial “gold rush” to publish on the boundary occurred after the formation of the working group on the PC-C boundary in the early 1970s; papers trailed off though the 1980s, and plummeted after the GSSP was ratified by the International Union of Geological Sciences (IUGS) in 1992 (Rowland and Corsetti, 2002). As our understanding of this interval of Earth history grows, we focus more on the “bigger picture” issues (e.g., evolution and diversification of the Metazoa), and focus less on the “boundary issues.” However, we will inevitably seek tie points with which to link fossils from siliciclastic sections to the geochemical and climate-change data from carbonate-dominated sections so that we can improve our understanding of this critical interval.

Mixed siliciclastic-carbonate successions become crucial in the search for stratigraphic tie points. Although a number of important lower Cambrian sections containing mixed siliciclastic-carbonate successions are

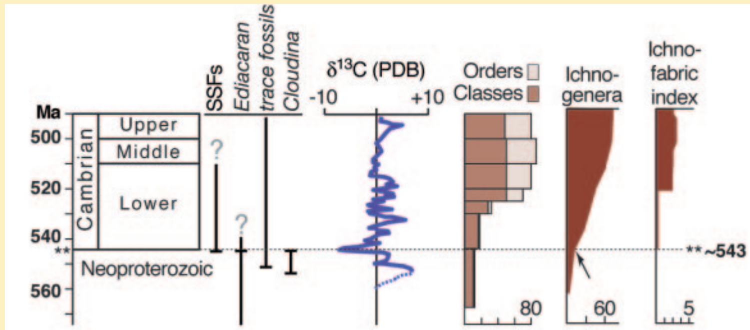

Figure 1. Summary of pertinent features associated with the PC-C transition (data compiled principally from Droser and Bottjer, 1988; Crimes, 1992; Droser et al., 1999, 2002; Knoll and Carroll, 1999; McIlroy and Logan, 1999), demonstrating the biospheric changes represented in this interval. SSFs: small shelly fossils. Arrow denotes the first appearance of vertically-oriented bioturbation.

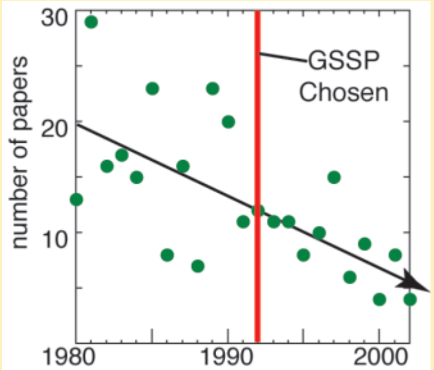

Figure 2. PC-C boundary publications through time (constructed by searching GEOREF for 'Precambrian-Cambrian,' 'Proterozoic-Cambrian,' and 'Neoproterozoic-Cambrian' in the title).

known (e.g., Mongolia, Brasier et al., 1996; Olenik uplift, Siberia, Knoll et al., 1995; Mackenzie Mountains, northwestern Canada, Narbonne and Aitken, 1995; southern Great Basin, Corsetti and Hagadorn, 2000), not all of these contain the PC-C boundary. Also, not all the sections contain an appropriate juxtaposition of carbonate- and trace-fossil-rich siliciclastic strata. Thick, relatively complete, well-exposed, and easily accessible successions in the southern Great Basin contain siliciclastic units and the boundary-marking fossil, *T. pedum*, in association with carbonate units recording a complete $\delta^{13}C$ chemostratigraphic profile (Corsetti and Kaufman, 1994; Corsetti et al., 2000; Corsetti and Hagadorn, 2000; Hagadorn and Waggoner, 2000). The southern Great Basin sections provide an excellent opportunity to compare results from a variety of stratigraphic approaches that have emerged as useful for correlating potential stage and series boundaries within the Cambrian (e.g., Geyer and Shergold, 2000).

GEOLOGIC BACKGROUND

From the time of Walcott (1908), the thick, superbly exposed, and highly fossiliferous Lower Cambrian strata from the southwestern United States (Figs. 3, 4, cover photo) have proved instrumental for understanding the Cambrian biotic explosion, and have even been suggested for a potential basal Cambrian stratotype (Cloud, 1973). Proterozoic—lower Paleozoic strata in the southern Great Basin were deposited on a thermally subsiding trailing margin created through rifting of the Laurentian craton in the Neoproterozoic (e.g., Stewart, 1966, 1970; Stewart and Suczek, 1977; Armin and Mayer, 1983; Bond et al., 1985).

Neoproterozoic-Cambrian strata in the southwestern United States thicken from southeast to northwest, and can be grouped into four distinct but interfingering successions: Craton, Craton Margin, Death Valley (proximal-shelf), and White-Inyo (proximal- to mid-shelf) successions (Stewart, 1970; Nelson, 1976, 1978; Mount et al., 1991; Corsetti and Hagadorn, 2000; Fedo and Cooper, 2001; Fig. 3). An erosional disconformity removed the PC-C boundary interval in the Craton and Craton Margin successions (Fedo and Cooper, 1990, 2001). This sequence boundary is traceable from the Craton Margin through the Death Valley succession to the White Inyo succession, and probably represents the "Sauk I" disconformity (Palmer, 1981). The boundary interval resides below this unconformity where incision was limited. The Death Valley succession records both appropriate trace fossil biostratigraphy and carbon isotope data (Corsetti and Hagadorn, 2000). However, the sections are relatively thin and the carbonates from which the $\delta^{13}C$ record was recovered are particularly thin. The White-Inyo succession is thickest, but, until now, has been considered poorly fossiliferous with respect to earliest Cambrian guide fossils (e.g., Signor and Mount, 1986). Trilobites, which are important guide fossils in Lower Cambrian sections in the Great Basin (e.g., Nelson, 1976; Palmer, 1981, 1998; Hollingsworth, 1999), make their first appearance well above this interval (Hollingsworth, 1999).

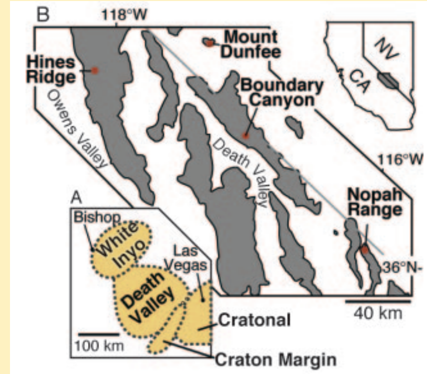

Figure 3. A. Map showing distribution of four interfingering facies successions that span the PC-C interval in the southern Great Basin; successions become progressively thicker to the northwest, the offshore direction (after Nelson, 1978; Corsetti et al., 2000; Fedo and Cooper, 2001). B. Shaded area represents the approximate PC-C outcrop belt in the southern Great Basin (after Stewart, 1970); the Mt. Dunfee section was offset to its present position by post-Cambrian transtensional faulting.

White-Inyo Succession

In ascending order, the White-Inyo Succession consists of the Wyman Formation, Reed Dolomite, Deep Spring Formation, Campito Formation, Poleta Formation, Harkless Formation, and the Mule Spring Limestone (Nelson, 1962; Fig. 4). The White-Inyo Mountains have been the focus of intense PC-C boundary study, but the paucity of earliest Cambrian fossils has been problematic (e.g., Cloud and Nelson, 1966; Taylor, 1966; Alpert, 1977;

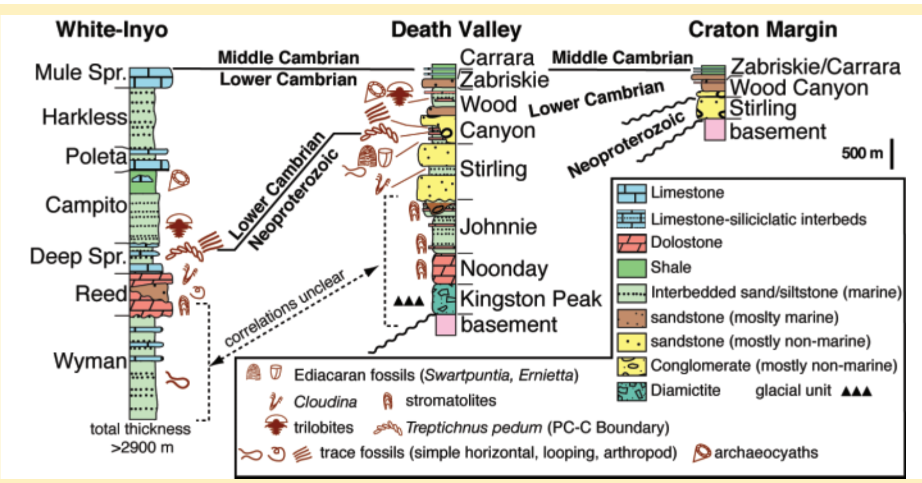

Figure 4. Generalized lithostratigraphic columns for the PC-C transition interval in the southern Great Basin (after Nelson, 1962, 1976; Stewart, 1970; Fedo and Cooper, 2001). Whereas the Lower Cambrian portion is well constrained and correlated between the intervals, correlations between Neoproterozoic strata are less well constrained. Base of the Sauk I Sequence removes some of the Neoproterozoic-Lower Cambrian interval in more proximal (Craton Margin) settings. The Death Valley succession contains a Neoproterozoic glacial-cap-carbonate succession, but no known correlatives exist in the White-Inyo succession. Fossil symbols represent first occurrences of key taxa.

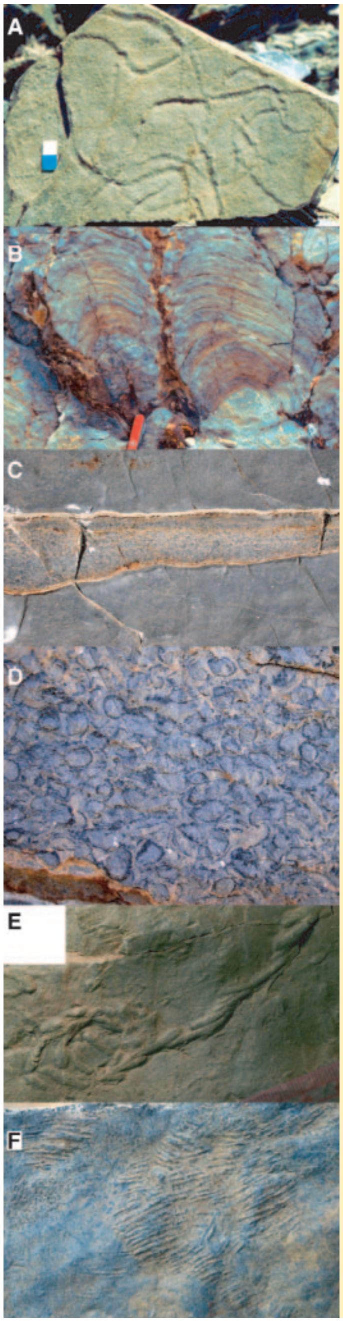

Figure 5: A. Helminthoidichnites on bed sole, Wyman Formation, south of Hines Ridge, Andrews Mountain, California (scale bar=1 cm increments). B. Domal stromatolites, Middle Deep Spring Formation, Mt. Dunfee, Nevada (knife ~8 cm long). C. Packstone lag bed composed mostly of Cloudina riemkeae (across middle of photo), Lower Deep Spring Formation, Mt. Dunfee, Nevada (field of view ~32 cm wide). D. Close-up of area shown in Fig. 5C showing C. riemkeae debris (field of view ~5 cm wide). E. Treptichnus pedum, Upper Deep Spring Formation, Andrews Mountain, California (field of view ~9 cm wide). F. Cruziana and Diplichnites, Upper Deep Spring Formation, Hines Ridge, California (field of view ~9 cm wide).

Nelson, 1976, 1978; Mount et al., 1983; Signor et al., 1983; Gevirtzman and Mount, 1986; Signor and Mount, 1986; Droser and Bottjer, 1988; Corsetti and Kaufman, 1994; Fritz, 1995; Hagadorn and Bottjer, 1999; Hagadorn et al., 2000). Because the PC-C boundary paradigm has changed over the last three decades, the inferred position of the boundary in the succession has changed as well. New fossil evidence (Fig. 5) is consistent with the position of the boundary. More comprehensive biostratigraphic information is contained in the cited references.

The Wyman Formation consists of interbedded mudrock, siltstone, and quartzite, with lensoidal oolitic, pisolitic, and oncolitic carbonate layers that increase in number upsection. Near Andrews Mountain in the Inyo Range, the formation exceeds 3000 meters in thickness (Nelson, 1962). The section primarily represents shallow marine deposition. The base of the formation is not exposed, so the nature of any underlying contact in not known (Nelson, 1962). The Reed Dolomite rests unconformably on the Wyman at most localities and is divided into three members: the Lower Member, Hines Tongue, and the Upper Member. The Lower Member is characterized by coarsely-crystalline pink dolostone with cross-bedded oolitic horizons, and minor domal stromatolite horizons. This suggests subtidal to intertidal marine deposition. The Hines Tongue is a southward-thickening siliciclastic unit consisting of hummocky-crossbedded sandstone and minor siltstone with minor carbonate interbeds. Thus suggests deposition below normal wave base but above storm wave base. The Hines Tongue is thickest in the Hines Ridge area of the Inyo Range, and thins dramatically to the north in the White Mountains and Esmeralda County, Nevada. The Upper Member is characterized by massive dolostones. Minor karstification is present at the contact with the overlying Deep Spring Formation at some localities.

The Deep Spring Formation is formally divided into the Lower, Middle, and Upper Members, each consisting of a siliciclastic-carbonate couplet (Gevirtzman and Mount, 1986; Mount et al., 1991). The siliciclastic half-cycle of each member contains green, ripple cross-laminated siltstones, and quartzites with hummocky cross-stratification, indicating deposition in relatively shallow water above storm wave base. The boundary between the siliciclastic and carbonate half-cycle is transitional at most localities. The carbonate half-cycle is commonly characterized by rhythmically interbedded carbonate wackestone and siliciclastic-rich siltstone, crossbedded oolite, and intraclastic grainstone; a high-energy, shallow-water depositional environment is indicated. The top of each carbonate half-cycle is commonly dolomitized, and often shows minor karstification; periodic emergence is indicated. Overall, sedimentary stacking patterns, sedimentary structures, and facies associations suggest that each member represents a shallowing-upward parasequence. Sharp disconformities divide the members (but see Mount et al., 1991). The Campito, Poleta, Harkless (and Saline Valley), and Mule Spring formations record similar shallow marine, mixed-siliciclastic carbonate strata (e.g., Moore, 1976; Mount and Bergk, 1998), and have similar sedimentary origins.

The White-Inyo succession contains a number of body and trace fossils. We have identified bed-parallel tubular trace fossils, including *Helminthoidichnites* and *Planolites* (Fig. 5A), in the Wyman Formation; we cannot falsify the hypothesis that some of these are body fossils. The Hines Tongue of the Reed Dolomite contains a depauperate suite of bed-parallel trace fossils, such as *Helminthoidichnites*, *Planolites*, and *Torrowangea*, and the Upper Member contains packstones of the body fossil *Cloudina*. Carbonates of the Lower and Middle Members of the Deep Spring Formation contain *Cloudina* (in skeletal lags and isolated occurrences; Figs. 5C, D). These fossils were initially identified as Cambrian small shelly fossils (Signor et al., 1983) but were subsequently reinterpreted as the Neoproterozoic *Cloudina* (Grant, 1990). It remains unclear whether all of the forms reinterpreted as *Cloudina* are, in fact, *Cloudina* or whether some are small shelly fossils. Siliciclastics of the Middle Member also contain rare examples of *Cloudina* and bed-parallel trace fossils, including *Planolites* and *Plagiogmus*. *Treptichnus pedum* (Fig. 5E), which delineates the PC-C boundary, occurs near the top of the Middle Member in the Mt. Dunfee area and at the base of the Upper Member in the White Mountains; in the latter case, *T. pedum* is associated with a moderately diverse ichnofossil assemblage (including *Cruziana* and *Rusophycus*, Fig. 5F). The Campito Formation contains trilobites characteristic of the *Fallotaspis* and *Nevadella* zones (see Hollingsworth, 1999), abundant trace fossils, and limited archaeocyathid bioherms (Nelson, 1976, 1978).

δ^{13}C CHEMOSTRATIGRAPHY

Faunal data are broadly useful for correlation across the PC-C interval in the southern Great Basin, but δ^{13}C chemostratigraphy provides another important technique for constraining intrabasinal and interre-

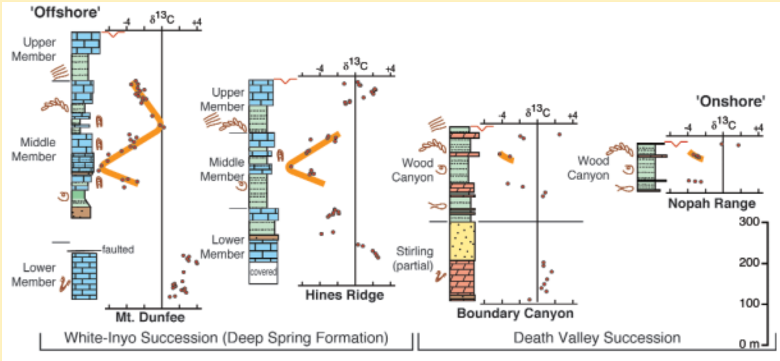

Figure 6. Integrated chemostratigraphy and biostratigraphy for the PC-C interval in the southern Great Basin (data from Corsetti and Kaufman, 1994; Corsetti and Hagadorn, 2000 and references therein; this study). The δ^{13}C record is most complete at the Mt. Dunfee section (the most offshore section), where T. pedum *occurs in the upper part of the Middle Deep Spring Formation. The d^{13}C record is progressively less complete towards the craton. Stromatolites occur in the White-Inyo succession in association with the δ^{13}C nadir.*

gional correlations. The Neoproterozoic-Cambrian δ^{13}C record has been relatively well characterized and includes many positive and negative excursions (e.g., Magaritz et al., 1991; Brasier et al., 1994, 1996; Strauss et al., 1992; Corsetti and Kaufman, 1994; Shields, 1999; Corsetti et al., 2000; Montanez et al., 2000) reflecting secular variation and recognizable globally. Although most δ^{13}C data are recovered from carbonate dominated successions, it is possible to analyze organic-rich siliciclastic successions for δ$^{13}C_{org}$. To be effective, this procedure requires that the analyzed section did not experience significant heating, and that rules out many sections from serious consideration (e.g., Strauss et al., 1992). Thus, robust δ^{13}C data from siliciclastic-dominated sections have remained elusive. A carbon isotope reference curve does not exist for the PC-C interval, but broadly similar chemostratigraphic patterns exist among many PC-C sections (Shields, 1999). Ignoring low amplitude variations, the major δ^{13}C trends include: 1, a latest Neoproterozoic major positive carbon isotope excursion (slightly older than 548 Ma; Grotzinger et al., 1995), associated with *Cloudina*, simple horizontal trace fossils, and Ediacaran-type fossils; followed by 2, a pronounced negative carbon isotope excursion nearly coincident with the PC- C boundary, at ca. 543-542 Ma (Bowring et al., 1993; Grotzinger et al., 1995). The precise position of the negative excursion with respect to the paleontologic marker of the boundary was unclear for a number of years, although it was commonly assumed that the negative excursion coincided with the boundary horizon. Corsetti and Hagadorn (2000) demonstrated that *T. pedum* does in fact occur within one negative- δ^{13}C shift of the boundary horizon in the Death Valley succession, and this potentially resolves the question.

The relative synchronicity of the first appearance of *T. pedum* and the negative δ^{13}C excursion in the southern Great Basin can be further tested by comparing samples from the thinner Death Valley succession to samples from the much thicker, carbonate-rich White-Inyo succession. High-resolution sampling for carbon isotope chemostratigraphy was conducted though the Deep Spring Formation at multiple sections across the basin, in concert with biostratigraphic sampling (Fig. 6). Most of the Lower Deep Spring Formation, where *Cloudina* is present, shows a positive isotopic excursion (to ~+4‰ PDB). The excursion is progressively omitted in the onshore direction, and reaches only ~+2‰ in more onshore sections. A negative excursion, commonly down to ~-5‰, is recorded from the top of the Lower Member through the middle of the Middle Member. Cloudina also occurs in this interval. The negative excursion is most pronounced in offshore sections, where isotopic compositions plummet to ~-7‰. Curiously, the isotopic nadir is associated with unusually abundant stromatolite and thrombolite development (Oliver and Rowland, 2002). At Mt. Dunfee, which represents the most offshore section, δ^{13}C values rise to near 0‰, then return to mildly negative values beneath the Middle-Upper Deep Spring contact. This excursion is missing from the other, less complete sections in the onshore direction. *T. pedum* occurs in association with this return to negative δ^{13}C values. The Upper Deep Spring Formation records a positive excursion to ~+2‰. Thus, the presence of *T. pedum* in association with the negative excursion is verified in the White-Inyo succession, and it provides a global tie-point useful for correlations between siliciclastic-dominated sections and carbonate-dominated sections.

In addition to providing a tool for chronostratigraphic work, secular variation in the δ^{13}C record can be used to address issues of basin scale. For example, if we use the δ^{13}C record as a chronostratigraphic tool, there is a progressive omission of the δ^{13}C record in the onshore direction. The most complete isotopic and stratigraphic records are present in the most offshore sections. This trend is not unexpected. However, previously it was not possible to determine the *magnitude* of stratal omission using available lithostratigraphic or biostratigraphic information.

GLOBAL IMPLICATIONS

Integrated biostratigraphic and chemostratigraphic information from the southern Great Basin demonstrate that the first occurrence of *T. pedum*, the trace fossil used to correlate the PC-C boundary, co-occurs with the ubiquitous negative carbon isotope excursion recorded in carbonate-dominated successions around the world. It is beyond the scope of this paper to correlate between all the carbonate- and siliciclastic-dominated section because endemism, hiatuses, and diagenesis complicate the global picture. Using the trace fossil and chemostratigraphic records from the Great Basin as a bridge, however, it appears that the first occurrence of small shelly fossils was relatively synchronous with the first appearance of *T. pedum*. If we ignore potential facies control on the first appearance of *T. pedum* in the Mt. Dunfee section, it could be argued that small shelly fossils just barely predate *T. pedum* because the first small shelly fossils appear in association with relatively negative δ^{13}C values (Knoll et al., 1995). Given the relatively small amount of stratigraphic uncertainty, the debate regarding the choice of trace fossils vs. small shelly fossils as the stratotype marker is, in our view, rendered moot. The PC-C transition is well calibrated (Bowring et al., 1993; Grotzinger et al., 1995), and the duration of the negative excursion is constrained to less than one million years. This implies that phosphatic biomineralization and vertically-oriented burrowing developed quickly and nearly synchronously (probably in less than one million years). Interestingly, the δ^{13}C and trace fossil biostratigraphic records from the Mt. Dunfee area closely match the hypothetical, composite reference section proposed by Shields (1999).

CONCLUSION

Stratigraphic sections in White-Inyo Mountains, California-Nevada, provide a well-exposed and easily accessible PC-C boundary interval through a mixed siliciclastic-carbonate succession. In this succession, $\delta^{13}C$ chemostratigraphy has been combined with biostratigraphy to provide well-constrained correlations, and these correlations have application for correlating between siliciclastic- and carbonate-dominated successions globally. The ubiquitous negative $\delta^{13}C$ excursion near the base of the Cambrian is confirmed to coincide with the first occurrence of *T. pedum* in multiple sections across the southern Great Basin. From the time of Walcott to the present day, the Neoproterozoic-Cambrian succession in the southwestern United States continues to provide important data on one of the most interesting intervals in Earth history.

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