Carbon isotope stratigraphy of Upper Cambrian (Steptoean Stage) sequences of the eastern Great Basin: Record of a global oceanographic event

Matthew R. Saltzman* **Bruce Runnegar** } *Department of Earth and Space Sciences, University of California, Los Angeles, California 90095*

Kyger C. Lohmann *Department of Geological Sciences, University of Michigan, Ann Arbor, Michigan 48104*

ABSTRACT

A large, global positive excursion in $\delta^{13}C$ (from ~0.5 to 4.5‰) during the Late Cambrian Pterocephaliid biomere/Steptoean Stage (*Aphelaspis-Elvinia* zones) is documented at high stratigraphic resolution in three sections in the eastern Great Basin. The excursion, which we refer to as the Steptoean Positive Carbon Isotope Excursion, or SPICE excursion, began coincident with a world-wide extinction event. The $\delta^{13}C$ data from the Great Basin reflect global seawater values in a wide range of lithologies, including oolitic grainstones, wackestones, thrombolitic boundstones, and flat-pebble conglomerates. We use a section at Shingle Pass in the southern Egan Range to divide the Pterocephaliid biomere into eight isotope steps that represent equal divisions of the $\delta^{13}C$ rise and fall (±4‰). This provides a basis for recognition of a revised chronostratigraphic framework for the Pterocephaliid biomere/Steptoean Stage.

Strata deposited during the beginning of the SPICE excursion record a major change in the pattern of sedimentation in the eastern Great Basin. This is reflected in a siliciclastic-carbonate transition at Shingle Pass, Nevada, and a carbonate-siliciclastic transition at the House Range and Lawson Cove sections in Utah. A regional siliciclastic influx recognized throughout the Great Basin occurs near the peak of the SPICE excursion. Carbon isotope analyses from cratonal sections in Wyoming provide independent evidence that a major sedimentary hiatus took place on the craton during the time of the SPICE excursion.

The correlated changes in $\delta^{13}C$, relative sea level, and the marine biota during the SPICE excursion provide remarkably detailed records of a major paleoceanographic event. We speculate that changes in sea level, climate, or tectonics may have triggered the SPICE excursion and coeval extinction event. Subsequent burial of organic carbon caused the increase in $\delta^{13}C$ and may have led to an interval of global cooling. The results of this study lend confidence to carbon-isotopic studies of pre-Mesozoic rocks.

INTRODUCTION

During the past 15 yr it has become increasingly clear that marine limestones and even dolostones may faithfully record the carbon isotopic composition of the ocean water in which they formed (Scholle and Arthur, 1980; Gao and Land, 1991; Wang et al., 1996). This is a surprising result, given the isotopic heterogeneity of modern carbonate sediments (Gonzalez and Lohmann, 1985), and it seems to indicate that early, low temperature diagenesis averages the isotopic composition of the originally heterogeneous components. Because this process involves diagenesis, many scientists have been skeptical of interpretations based on carbon isotope stratigraphy, particularly when used as indicators of time in otherwise undatable Precambrian sedimentary sequences (e.g., Kaufman and Knoll, 1995). In order to address this issue and test whether carbon isotope excursions are useful for international correlation, we decided to investigate the carbon isotope stratigraphy of the oldest well-dated carbonates of the richly fossiliferous (Phanerozoic) part of the geological record. These are horizons in the Middle and Late Cambrian that have been correlated globally using agnostid trilobites.

The two horizons chosen initially for investigation are different in character. The older horizon represents the evolutionary transition from one species of the trilobite *Ptychagnostus* to another species of the same genus (Rowell et al., 1982). Faunal turnover associated with this transition is minor, and, not surprisingly, the carbon isotope record is unremarkable. However, the younger horizon is a global mass extinction of trilobites first detected in northern China (Walcott, 1913) and subsequently documented across Australia (Öpik, 1966) and North America (Palmer, 1965b, 1984). It corresponds to the Steptoean, Idamean and Changshanian stages of North America, Australia, and China, respectively. It is also equivalent to the base of one of Palmer's (1965a) biomeres—the abrupt replacement of an established assemblage of epicontinental trilobites by another, less diverse offshore assemblage.

Data accumulated by M. D. Brasier, R. L. Ripperdan, and ourselves have shown that a major positive excursion (+4‰) in the carbon isotope ratio of carbonate rocks begins at this mass extinction horizon (Saltzman et al., 1995a, 1998). We refer to this excursion as the Steptoean Positive Carbon Isotope Excursion (SPICE), using the stage nomenclature of Ludvigsen and Westrop (1985). The SPICE excursion is known to occur in Kazakhstan, China, Australia and western North America and has also recently been found in eastern North America (Glumac and Walker, 1996). It is among the largest positive isotope excursions of the past 500 m. y. and is one of the first pre-Cenozoic events to be shown to be truly global in distribution. Similar events in the Mesozoic have been well-documented in the North Atlantic region (see Schlanger et al., 1987). Here, we present the results of detailed investigations of the SPICE excursion in western North America. The primary goal is to utilize the carbon isotope curves to improve chronostratigraphic correlation among the three stratigraphic sections analyzed from the eastern Great Basin of the United States (Fig. 1).

*e-mail: saltzman@ess.ucla.edu

Data Repository item 9804 contains additional material related to this article.

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We also examine the paleoenvironmental significance of the carbon cycle perturbation indicated by the SPICE excursion by addressing its correlation with changes in relative sea level and the biota.

GEOLOGIC BACKGROUND AND STRATIGRAPHIC FRAMEWORK

The time of the SPICE excursion in North America—the Pterocephaliid biomere, using Palmer's nomenclature (1965a) or the Steptoean Stage sensu Ludvigsen and Westrop (1985)—is bounded by two sharply defined mass extinction horizons. These represent nearly isochronous surfaces (Palmer, 1979) and provide the independent evidence for correlation of our isotope curves in the eastern Great Basin. The Pterocephaliid biomere/Steptoean Stage is exceptionally complete in the Great Basin region where Palmer (1965b) recognized five trilobite assemblage zones (including, from oldest to youngest, the *Aphelaspis*, *Dicanthopyge*, *Prehousia*, *Dunderbergia*, and *Elvinia* zones) and to document their evolutionary continuity.

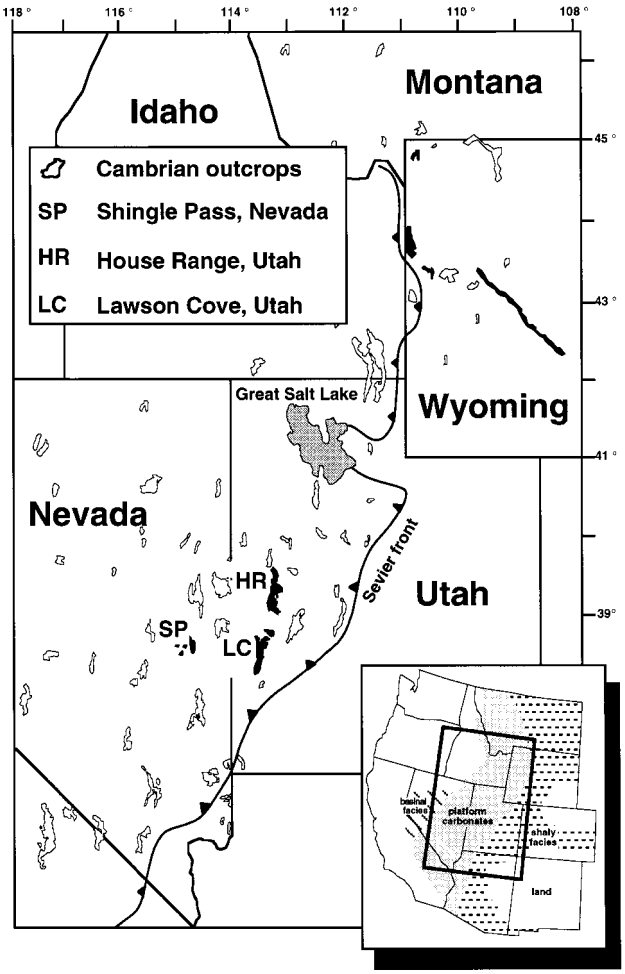

Figure 1. Map of the study area in the western United States showing outlines of major Cambrian outcrops and trace of the Sevier orogenic belt (after Palmer, 1971). The focus of this paper is on the stratigraphy of the three outcrop areas indicated in black near the Utah-Nevada border. Additional outcrops in black in Wyoming were sampled for isotopes.

Sedimentation in the Great Basin during the Pterocephaliid biomere/Steptoean Stage occurred as part of an extensive drift-stage carbonate platform, which characterized the Cordilleran miogeocline beginning in Middle Cambrian time (Armstrong, 1968; Palmer, 1971; Stewart and Poole, 1974; Levy and Christie-Blick, 1991). Carbonate platform sedimentation in the region was primarily influenced by short-term changes in eustatic sea level, superimposed on the long-term Sauk transgression (Sloss, 1963), and a structural feature known as the House Range embayment, which affected local topography and subsidence rates (Rees, 1986). Climate may also have affected sedimentation by controlling siliciclastic influx and the health of the carbonate platform (Cowan and James, 1993). These factors resulted in considerable lateral and vertical facies changes on the carbonate platform and a highly variable siliciclastic component (Palmer, 1971; Koepnick, 1976; Rowell and Brady, 1976; Lohmann, 1976).

The three sections analyzed for carbon isotopes (Fig. 1) were chosen to be representative of this variability in facies successions. Furthermore, they were chosen because of the availability of detailed biostratigraphic and lithostratigraphic frameworks. The three sections (and the source of these frameworks) include: (1) the House Range (Hintze and Palmer, 1976; Rees et al., 1976; Sims, 1985); (2) Lawson Cove (Hintze and Palmer, 1976; Lilley, 1976); and (3) Shingle Pass (Kellogg, 1963; Lilley, 1976; Rowell and Brady, 1976). Although the boundaries of the Pterocephaliid biomere/Steptoean Stage can be located within a few meters of section at these localities, the boundaries between the trilobite zones within the biomere/stage are often poorly constrained because of limits in the spacing of collections and the preservation state of the material collected. The carbon-isotope stratigraphy outlined below thus represents a basis for improved confidence in time correlation within the biomere/stage, not only among the three sections in the eastern Great Basin we have focused on, but among other sections through this time interval that can be analyzed for carbon isotopes.

GEOCHEMICAL PROCEDURES: DEALING WITH DIAGENESIS

Several studies have addressed the extent to which diagenesis may alter the original marine isotopic ratios recorded by carbonate rocks. Magaritz (1983) and Banner and Hanson (1990) showed that extremely high water:rock ratios are necessary to significantly alter primary $\delta^{13}C$ values. Other investigations have sought to evaluate the preservation and correlation of secular trends. Scholle and Arthur (1980) demonstrated that major fluctuations in the bulk $\delta^{13}C$ ratios of Cretaceous limestones were visible from one section to another despite differences in degree of diagenetic alteration. Similarly, Joachimski and Buggisch (1993) and Saltzman et al. (1995b) presented evidence for the preservation of trends in $\delta^{13}C$ values from widely separated Upper Devonian and Upper Cambrian stratigraphic sections, respectively.

In working with Proterozoic carbonate successions that lack independent evidence for correlation, Kaufman et al. (1991) and Derry et al. (1992) developed empirical indicators for the effects of diagenesis on primary $\delta^{13}C$ values, including screens based on isotopic, elemental and petrographic analysis, and total organic carbon content. These and subsequent investigations (Narbonne et al., 1994; Pelechaty et al., 1996) indicated that the vast majority of Proterozoic carbonates did not show evidence for alteration sufficient to greatly change the original $\delta^{13}C$ values. They also presented evidence indicating that the effects of sedimentary facies and lateral variations on the C-isotope ratios were minimal. In fact, Narbonne et al. (1994; p. 1286) went as far as to suggest that "the best way to be confident that stratigraphic variation seen in individual sections corresponds to secular changes in the world's ocean is to demonstrate that the same stratigraphic variations occur in different sections that can be shown by *independent* means to be of comparable age" (our italics).

Based on the conclusions of the aforementioned studies and our own experiences with Cambrian carbonates, our approach has not involved systematic geochemical screening of samples for diagenesis or rigorous characterization of intrasample or bed variability. We used a microscope-mounted drill assembly in order to easily obtain carbonate powder from petrographically least altered areas of polished thin-section billets. Homogeneous micrites were the preferred component, although many of the samples (~1 mg of powder) isolated had a coarser-grained component that was recrystallized to various degrees. This reflects the remarkably wide range of carbonate lithologies to be sampled, including oolitic packstones, skeletal wackestones, thrombolitic boundstones, and flat-pebble conglomerates. Although it is likely that some of these samples may not be representative of the bed from which they came or the seawater from which they formed, the independent fossil evidence for correlation of the secular trends suggests that most are. We predict that future studies designed to better characterize signal quality will result in further support of this conclusion.

Powders from drilling (~1 mg) were roasted for 1 h at 380 °C and reacted with 100% H_3PO_4 at 72 °C in a Finnigan Kiel extraction system coupled directly to a Finnigan MAT 251 mass spectrometer. NBS-18 and NBS-19 were run on a daily basis to maintain calibration and monitor analytical precision. All stable isotope ratios ($\delta^{13}C$ and $\delta^{18}O$) are normalized to V-PDB scale based upon a least squares fit of observed to predicted values. Values ($\delta^{13}C$) for NBS-18 = –5.0 and NBS-19 = 1.95‰.

CARBON ISOTOPE STRATIGRAPHY

The results of approximately 200 carbon isotope analyses performed on limestones from the three sections in the Great Basin are shown in Figure 2 and tabulated in Table 1 (Shingle Pass) and Tables DR 1 and 2[1] (Lawson Cove and House Range). The paired $\delta^{18}O$ values generally do not covary with $\delta^{13}C$ values and show no obvious stratigraphic trends. We use the $\delta^{13}C$ stratigraphic profile from the Shingle Pass section to divide the Pterocephaliid biomere/Steptoean Stage into eight isotope steps that provide the basis for improved chronostratigraphic correlation among the three sections (Figs. 3 and 4). We have chosen Shingle Pass as the reference section because sedimentation rates are particularly high and limestone deposition is relatively continuous through all of the trilobite zones of the Pterocephaliid biomere/Steptoean Stage.

The eight isotope steps illustrated in Figure 4 represent arbitrary 1‰ divisions of the SPICE excursion based on the three-point running average at Shingle Pass. The base of isotope step 1 corresponds to the beginning of the SPICE excursion, which is defined by the transition from $\delta^{13}C$ values fluctuating (±0.5‰) about a mean of 0.5‰ to values rising monotonically above that mean. Sections that are complete through the Pterocephaliid biomere/Steptoean Stage should thus record all eight isotope steps, and strata deposited during each individual step are assumed to be time-equivalent. What follows is a description and environmental interpretation of the sediments that accumulated during the SPICE excursion at Shingle Pass, Lawson Cove, and the House Range (Fig. 5). The effects of compaction are not considered. This step-by-step description provides a basis for discussion on the correlation of significant lithologic boundaries with important biostratigraphic and chemostratigraphic boundaries. In addition, it reveals the remarkable range of carbonate environments and lithologies that record global seawater $\delta^{13}C$ ratios.

[1]GSA Data Repository item 9804, tables of stable isotope data, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301; e-mail: editing@geosociety.org.

TABLE 1. STABLE ISOTOPE DATA FROM SHINGLE PASS, NEVADA

Height (m)	$\delta^{13}C$ (PDB)	$\delta^{18}O$ (PDB)	Height (m)	$\delta^{13}C$ (PDB)	$\delta^{18}O$ (PDB)	Height (m)	$\delta^{13}C$ (PDB)	$\delta^{18}O$ (PDB)
1.4	0.82	–12.46	112.9	0.67	–9.27	245.0	4.00	–9.36
2.8	0.57	–12.96	113.4	0.86	–9.43	252.0	4.65	–8.69
4.2	0.49	–9.60	114.8	0.95	–10.27	263.2	4.23	–10.75
5.6	0.24	–10.39	116.2	1.13	–10.01	273.0	4.04	–9.81
19.6	0.32	–9.54	117.6	1.05	–10.50	274.4	4.35	–9.12
28.0	0.44	–9.40	119.0	1.28	–10.26	277.2	4.42	–9.24
30.8	0.30	–9.20	123.2	1.11	–11.74	278.9	4.02	–9.42
35.0	0.66	–9.82	126.0	1.13	–10.85	279.0	4.20	–9.18
39.2	0.02	–9.49	130.2	1.41	–10.27	280.0	4.19	–9.12
40.6	0.45	–9.68	134.4	1.42	–10.23	281.4	3.86	–9.62
42.0	0.18	–9.09	137.2	1.55	–9.87	282.8	3.99	–9.96
43.4	0.10	–9.14	138.6	1.52	–9.25	285.6	4.30	–9.10
44.8	–0.41	–9.65	142.8	1.59	–10.20	286.3	4.52	–9.43
58.8	0.69	–9.70	144.2	1.98	–9.81	287.0	3.95	–10.54
68.6	0.34	–13.08	145.6	1.71	–10.12	287.7	4.34	–9.09
70.0	0.42	–8.98	147.0	2.18	–9.78	288.4	4.10	–9.03
78.4	0.47	–9.08	148.4	2.12	–10.41	289.8	3.17	–13.13
79.8	0.50	–9.51	149.8	1.86	–10.30	295.1	4.65	–9.33
84.0	0.37	–9.74	151.2	1.19	–7.58	295.4	4.54	–9.69
86.1	0.28	–9.83	155.1	1.92	–9.39	296.2	4.42	–9.21
86.8	0.08	–9.66	155.4	1.93	–9.10	296.8	4.16	–9.21
93.8	0.11	–10.25	156.8	1.64	–10.59	298.2	3.61	–9.24
95.2	0.77	–9.88	159.6	2.76	–9.39	299.0	3.85	–8.97
98.0	0.31	–9.16	161.0	2.33	–9.57	299.6	4.09	–8.95
98.4	0.47	–9.78	162.4	2.15	–9.85	301.0	3.97	–8.80
99.4	0.27	–9.40	163.8	2.33	–9.76	302.4	3.33	–9.36
100.5	0.41	–9.29	166.6	2.33	–9.14	304.5	2.96	–10.16
100.8	0.53	–9.55	169.4	2.38	–9.84	304.9	3.27	–9.63
101.0	0.20	–9.12	172.2	2.25	–9.05	306.6	3.11	–9.88
103.0	0.73	–9.42	173.6	2.79	–13.14	308.0	3.70	–9.46
103.7	0.46	–9.31	176.4	2.59	–10.47	313.6	3.06	–10.66
104.2	0.26	–9.23	177.8	2.80	–9.38	315.0	2.64	–9.64
104.7	0.49	–9.28	180.6	3.13	–11.13	316.4	2.60	–9.91
105.0	0.25	–9.49	183.4	3.18	–8.25	320.6	2.37	–9.31
105.1	0.53	–9.14	186.2	4.19	–11.83	322.3	3.59	–11.44
105.4	0.37	–9.70	189.0	3.11	–10.85	323.0	2.40	–7.75
106.0	0.51	–9.52	191.8	3.23	–10.60	324.8	1.66	–9.15
106.4	0.42	–10.22	193.2	2.91	–14.34	341.6	1.11	–9.45
107.0	0.98	–9.73	208.6	3.50	–12.58	348.6	0.68	–9.52
107.4	0.97	–11.03	212.8	3.73	–11.69	355.6	0.67	–8.97
107.5	1.05	–9.20	214.2	4.02	–9.75	364.7	0.71	–9.22
109.0	0.61	–9.17	229.6	4.17	–9.92	376.6	0.72	–9.04
110.0	1.57	–7.77	238.0	3.84	–8.72			

Note: Base of the Pterocephaliid biomere is at 106.8 m in the section.

Shingle Pass

At Shingle Pass in the southern Egan Range, the beginning of isotope step 1 coincides precisely with the mass extinction horizon that defines the base of the Pterocephaliid biomere/Steptoean Stage (Palmer, 1965a, 1979). The paleontological basis for location of this extinction is the identification of trilobite specimens of *Coosella perplexa* and a coquina of *Glaphyraspis* by Thomas (1993). The lower 7 m of step 1 are predominantly thin-bedded calcareous shales, wackestones, and flat-pebble conglomerates. This lithofacies association, similar to the underlying ~250 m of the Emigrant Springs Formation (Fig. 2; Kellogg, 1963), is interpreted to reflect quiet-water deposition in an intrashelf basin that was episodically scoured by storms (Markello and Read, 1982). The overlying 33 m of step 1 consist of nonfossiliferous cross-bedded peloidal grainstone and indicate a transition to a restricted, lagoonal environment. The contact between the recessive shaly unit and the cliff-forming peloidal grainstone unit is sharp and locally erosional.

Sediments deposited during isotope steps 2 and 3 consist of a succession of fenestral mudstone, oncolitic/oolitic grainstone, and bimodally cross-stratified echinodermal grainstones (Fig. 6D). This sequence is interpreted to reflect a gradual transition from restricted intertidal to supratidal depths to an open marine shoal-water complex of ooid sands and skeletal banks.

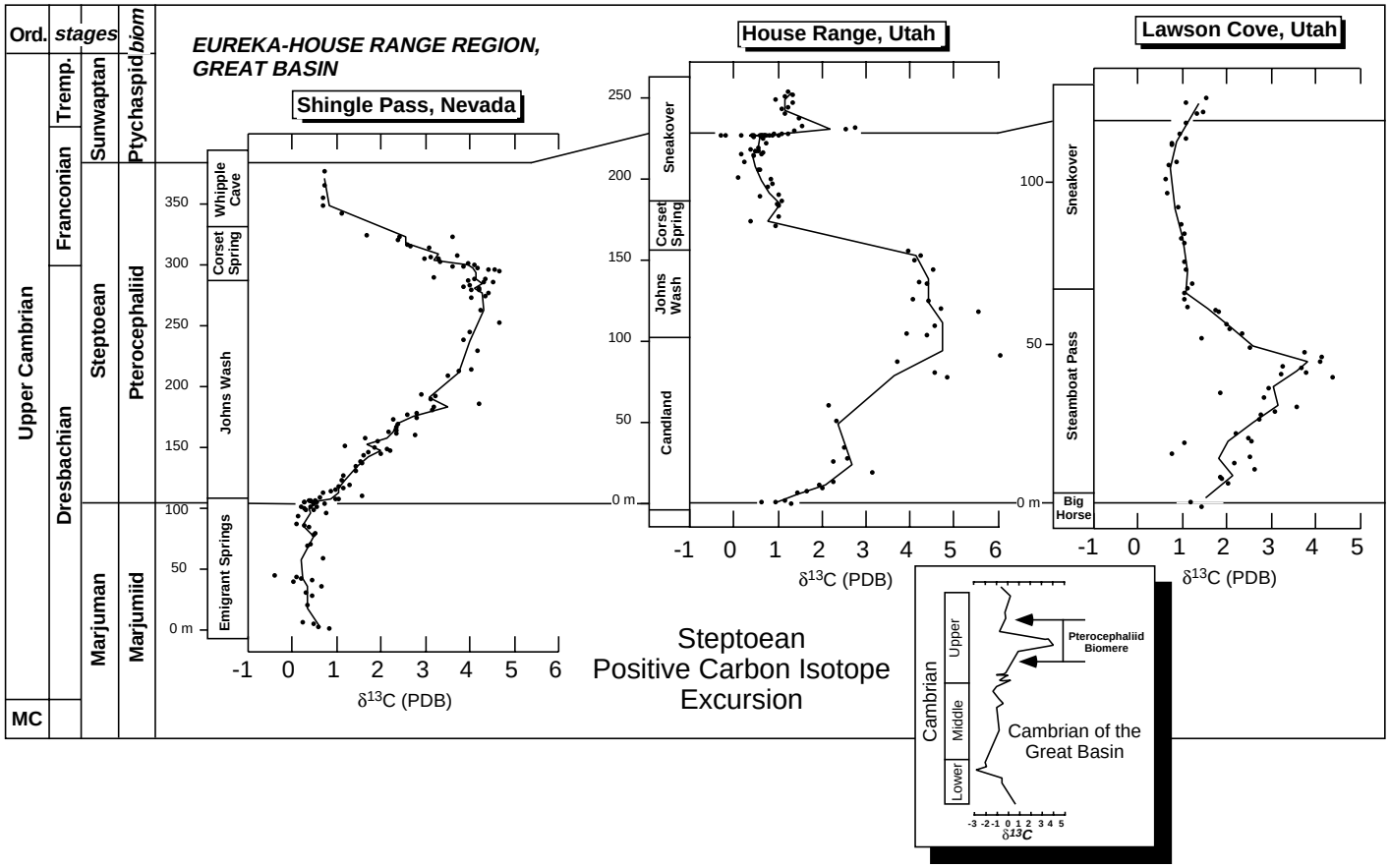

Figure 2. Stratigraphic $\delta^{13}C$ data (see Tables 1, DR1, and DR2 [see footnote 1 in text]) for three sections in the Eureka–House Range region of the Great Basin (see Fig. 1 for localities). Solid line representing a three-point running average is drawn through raw data points (solid circles). Biomere boundaries (Palmer, 1965a) and dual stage nomenclature (Ludvigsen and Westrop, 1985) are indicated. Formation boundaries are shown for each section (Kellogg, 1963; Palmer, 1965b; Hintze and Palmer, 1976). Inset is composite isotope curve for the Cambrian of the Great Basin from Brasier (1993) showing the presence of the SPICE excursion within the Pterocephaliid biomere.

The shoal-water complex may also be interbedded with thin (1 m) units of burrowed wackestone and thrombolitic boundstone as parts of upward-shallowing carbonate cycles. These quieter water facies indicate deposition in topographic depressions between migrating shoals.

The shelf cycles continue into isotope step 4, although these cycles gradually display thicker basal units of burrowed wackestone, thinner oolitic/skeletal units, and caps of peritidal to supratidal fenestral mudstone. These differences, compared with steps 2 and 3, seem to reflect a position on the shelf further from the migrating shoal. The upper 15 m of step 4 consist of alternating (meter-scale) units of burrowed wackestone and poorly exposed shale. This marks the first siliciclastic influx since the basal beds of step 1 and indicates a deepening of environments relative to the underlying shoal.

The succession deposited during steps 5 and 6 includes meter-scale cycles of shale, wackestone, and minor oolitic wackestone, which suggests renewed proximity to the shoal complex. Shale becomes dominant toward the top of the succession forming a largely covered slope and indicating an abrupt transition back to a deeper water siliciclastic setting. During step 7, these shales are abruptly overlain by resistant beds of cherty wackestone (which exhibit severe brecciation apparently related to local faulting). The cherty wackestone facies reflects a deep subtidal environment below storm-wave base (Brady and Rowell, 1976) and marks a significant change in the pattern of sedimentation at Shingle Pass that continues through the overlying time of isotope step 8.

Isotope step 8 is defined by the return of $\delta^{13}C$ to ratios <1‰, which marks the reestablishment of the Upper Cambrian baseline values that preceded isotope step 1. The top of step 8 is defined by a positive shift in $\delta^{13}C$ to values >1‰, which begins at the extinction horizon at the top of the Pterocephaliid biomere/Steptoean Stage (Saltzman et al., 1995b). This extinction horizon is not known precisely at Shingle Pass (Kellogg, 1963; Palmer, 1965b), and the top of isotope step 8 was not located.

Lawson Cove

At the Lawson Cove section in the southern House Range, the base of isotope step 1 is not precisely located. Thomas (1993) narrowed the base of the Pterocephaliid biomere/Steptoean Stage to within six m of section. We chose the base of this 6 m interval as the base of step 1. Sediments deposited during step 1 consist of 5 m of massive mudstone, algal boundstone, and peloidal packstone exhibiting fenestral fabric (Fig. 6A), overlain by 2 m of parallel to wavy laminated (mm scale) cryptalgal laminite (Fig. 6B) with minor intraclast breccia. This lithofacies association, which also character-

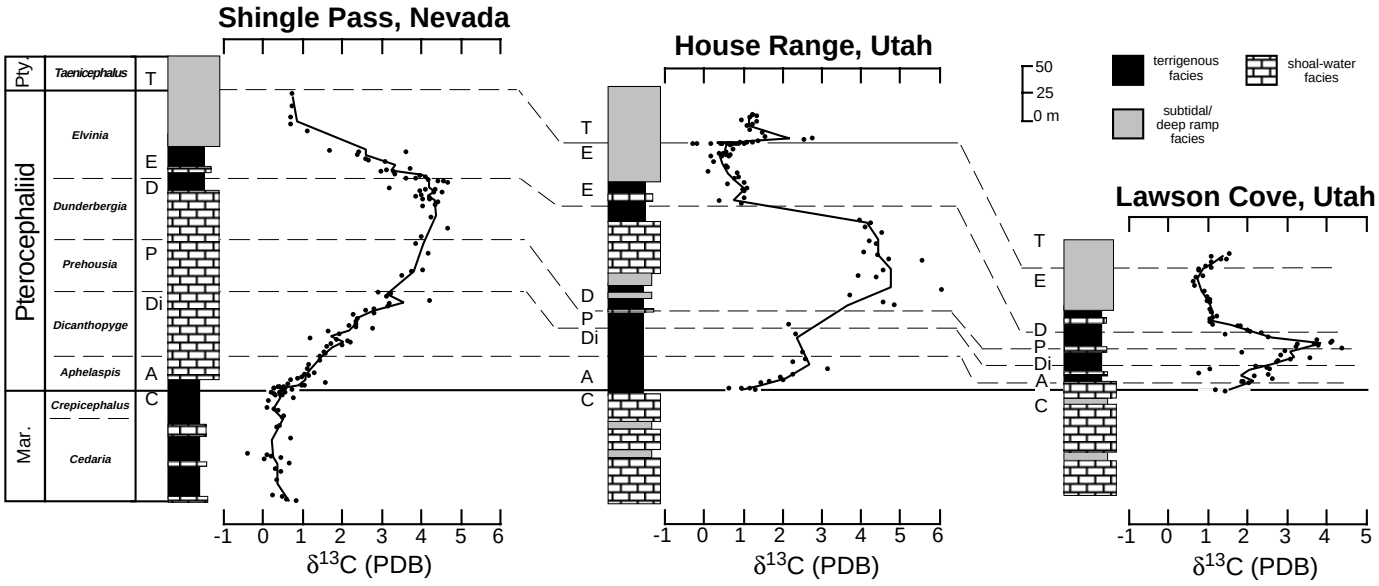

Figure 3. Stratigraphic $\delta^{13}C$ data (see Fig. 2, Tables 1, DR1, and DR2 [see footnote 1 in text]) for the House Range, Lawson Cove, and Shingle Pass sections showing approximate boundaries (dashed lines) of five trilobite zones of the Pterocephaliid biomere/Steptoean Stage. Critical fossil collections for biostratigraphic age assignments are indicated by capital letters next to the generalized stratigraphic columns. Biostratigraphic data from Palmer (1965b), Hintze and Palmer (1976), and Thomas (1993).

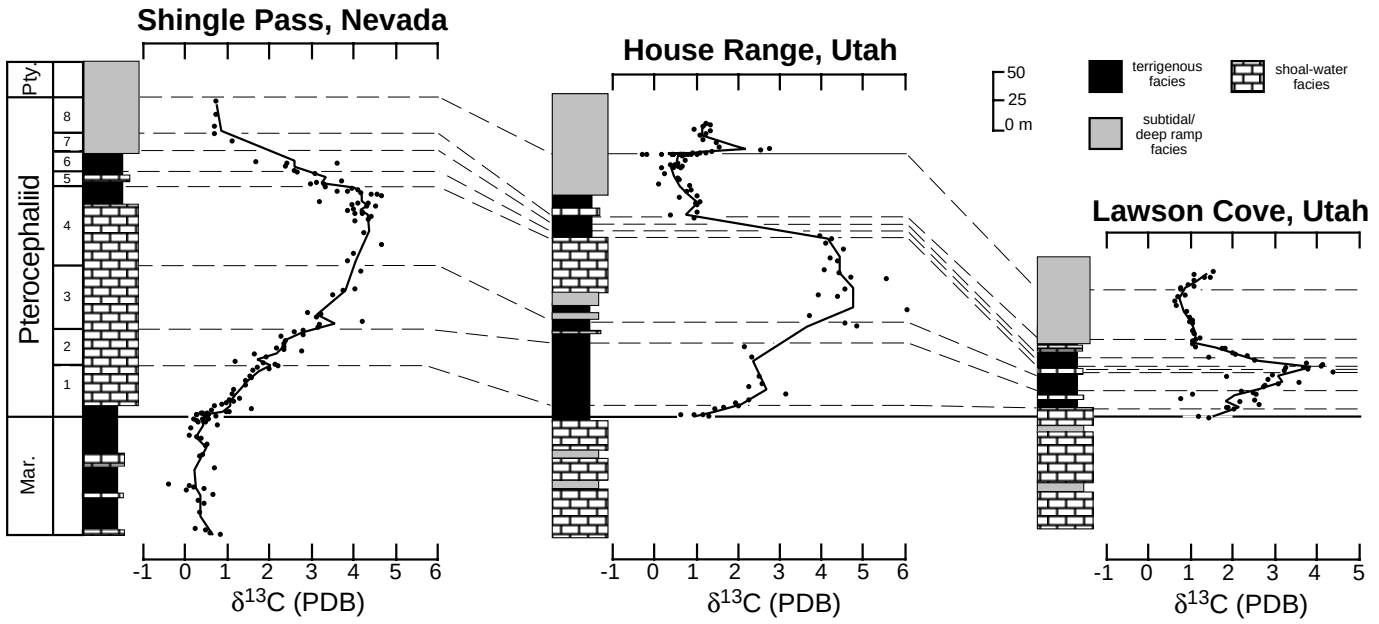

Figure 4. Stratigraphic $\delta^{13}C$ data (see Fig. 2, Tables 1, DR1, and DR2 [see footnote 1 in text]) showing boundaries (dashed lines) of eight isotope steps used in correlation among the three sections (compare with Fig. 3). Isotope steps were defined as arbitrary 1‰ divisions of the SPICE excursion based on the three-point running average (solid line) for the Shingle Pass section.

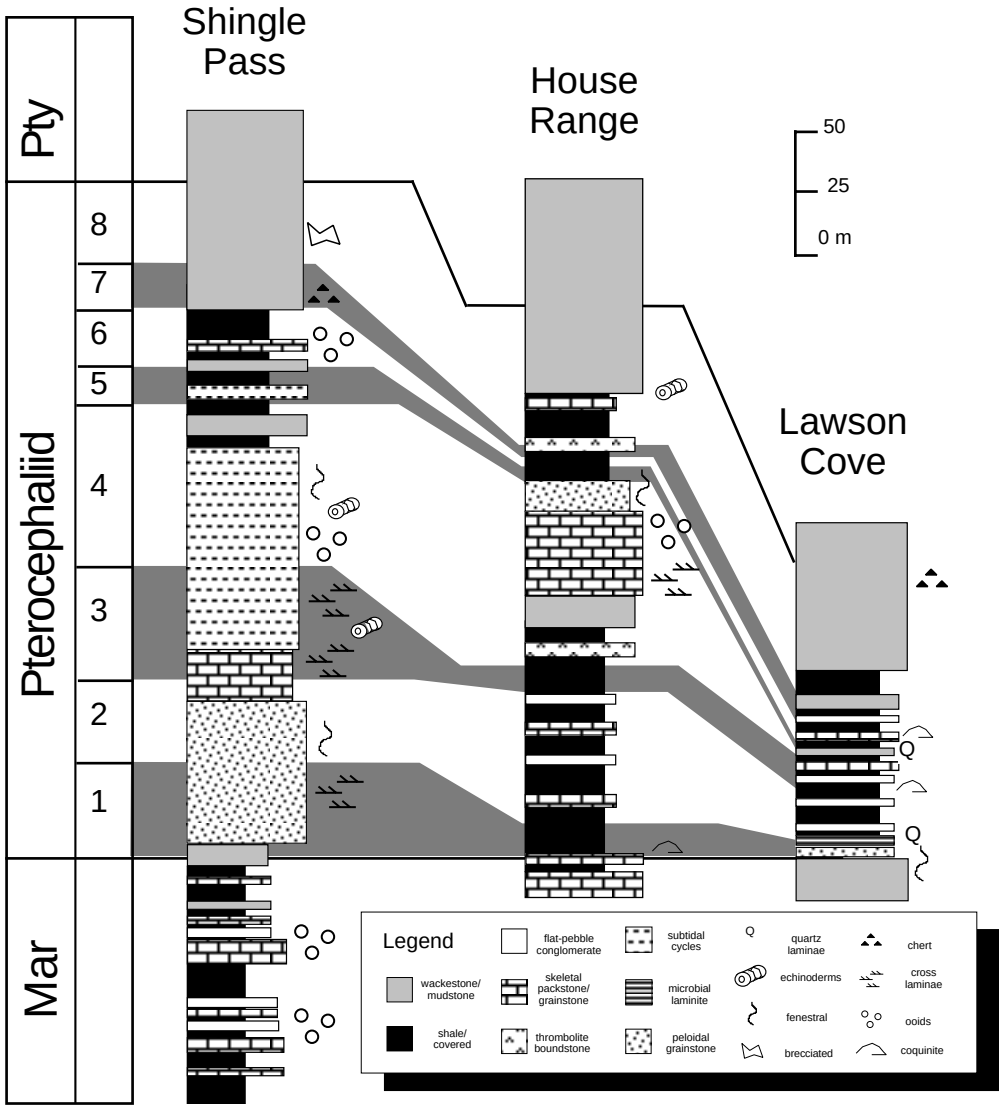

Figure 5. Stratigraphic columns showing important lithologic divisions as well as lateral thickness trends (shading) of the eight isotope steps defined in Figure 4.

izes the underlying ~200 m of Big Horse Limestone (Hintze and Palmer, 1976), indicates platform deposition in intertidal to supratidal environments (Rees et al., 1976; Westrop, 1989; Cowan and James, 1993; Osleger and Montanez, 1996). Thin beds of poorly exposed shale abruptly overlie the cryptalgal laminite at the top of step 1 and indicate a transition to a deeper water, low energy setting.

Sediments deposited during step 2 consist mainly of a shale and flat-pebble conglomerate (Fig. 6E) facies association. A 0.5-m-thick bed of cross-bedded sandy limestone occurs low in the succession and indicates a brief shallowing of water depths. This shaly storm-dominated intrashelf basin environment characterizes deposition at Lawson Cove through the middle of isotope step 7. During step 3, significant amounts of trilobite packstone are interbedded with the shale and flat-pebble conglomerate, possibly reflecting improved circulation within the intrashelf basin environment. Steps 4 and 5 include resistant beds of burrowed wackestone exhibiting abundant millimeter-scale laminations of angular quartz silt. The silt is likely eolian in origin and may be concentrated as a result of a decrease in the relative proportion of carbonate and shale. Exceptional exposures of olive-green shale occur in the middle part of step 6.

The upper 7 m of isotope step 7 and all of step 8 consist almost entirely of light gray wackestone/lime mudstone ledges. This homogenous carbonate unit is similar to that deposited during this time at Shingle Pass and represents the subtidal blanket carbonate of Brady and Rowell (1976). These authors interpreted this widespread facies association to reflect deposition below normal wave base on a nearly featureless shelf. The precise location of the extinction horizon marking the top of the Pterocephaliid biomere/Steptoean Stage is not known at Lawson Cove, but we were able to capture the beginning of the positive excursion to values >1‰ that marks the top of step 8 (Fig. 4).

Little Horse Canyon

At the Little Horse Canyon section in the northern House Range, sampling was not sufficient to reveal the base of isotope step 1 ($\delta^{13}C$ ratios <0.5‰). Palmer (1984) has precisely located the base of the Pterocephaliid biomere/Steptoean Stage, and we used this extinction horizon as the base of step 1 for our study (Fig. 4). Sediments deposited during step 1 consist of poorly exposed shale and calcareous shale with minor interbeds of skeletal

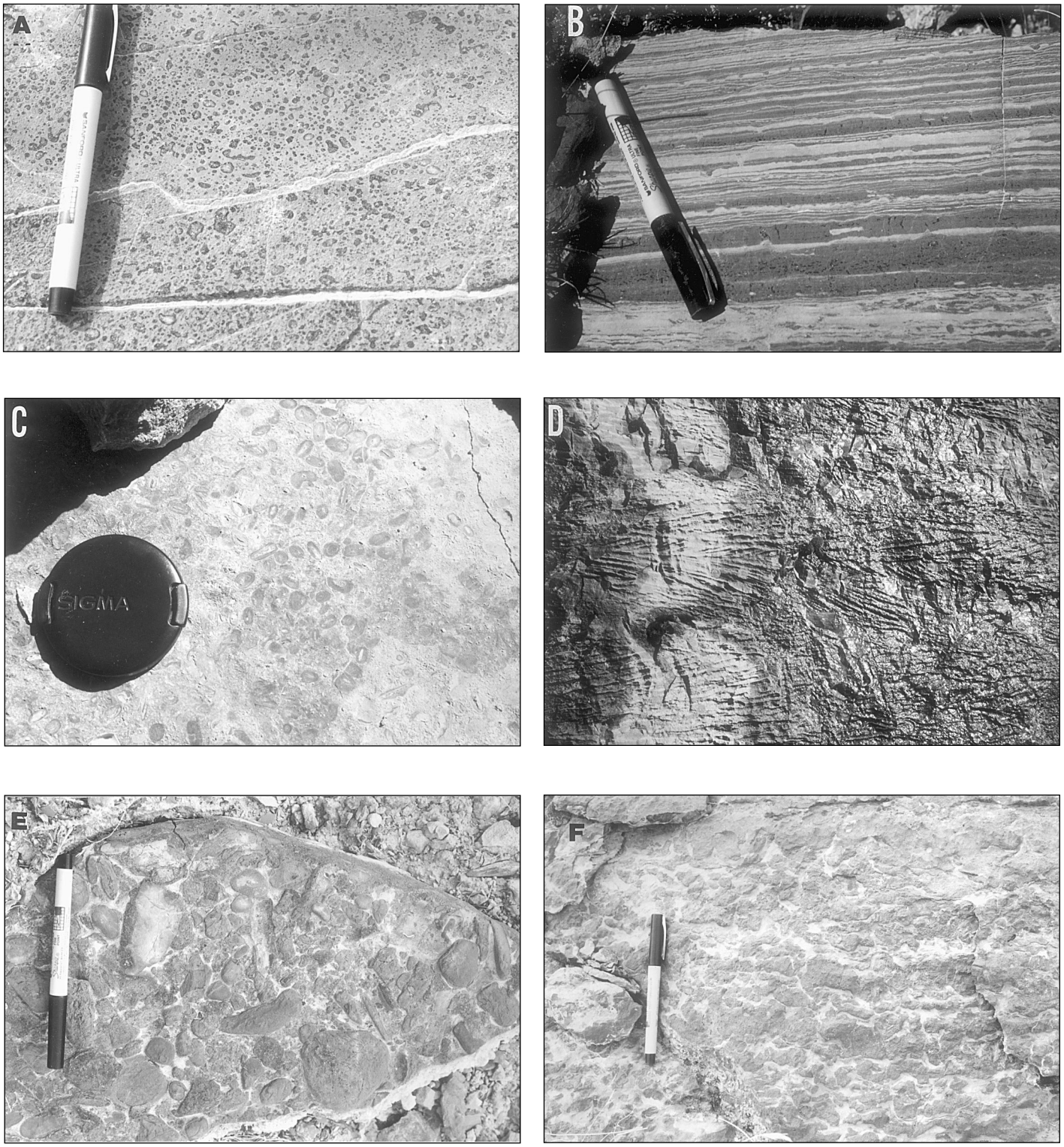

Figure 6. Field photographs of Upper Cambrian lithofacies. (A) Bedding plane view of lime mudstone facies exhibiting fenestral fabric. Lawson Cove section. (B) Cryptalgal laminite facies. Light layers are dolomite; dark layers, limestone. Lawson Cove section. (C) Bedding plane view of oncolite facies. Lens cap is 4.5 cm across. House Range section. (D) Bimodal planar cross-bedding in a peloidal grainstone unit. Field of view is 0.5 m across. Shingle Pass section. (E) Bedding plane view of flat-pebble conglomerate facies. Lawson Cove section. (F) Burrowed peloidal wackestone/packstone facies. Lawson Cove section.

or oolitic/oncolitic packstone (Fig. 6C). The shaly association marks an abrupt change from the underlying ~200 m of nearly pure carbonates of the Big Horse Limestone (Hintze and Palmer, 1976) and is interpreted to mark a transition to a deeper subtidal setting nearby a migrating shoal complex.

This shaly environment continues through steps 2, 3, and the early part of step 4, resulting in a largely covered slope with minor ledges of trilobite wackestone. Sediments deposited during step 4 record a large-scale upward shallowing unit that has been studied in detail by Rees et al. (1976). The succession includes massive, burrowed wackestone overlain by cross-bedded oolitic grainstone and capped by beds of lime mudstone and peloidal packstone exhibiting fenestral fabric. Rees et al. (1976) suggested the succession reflects the oceanward migration of a high-energy shoal environment, followed by migration of the lagoonal and tidal flat environments, which had flanked the shoal on its landward side. The overlying olive-green shale, burrowed wackestone, skeletal packstone, and flat-pebble conglomerate of steps 5, 6, 7, and 8 indicate a significant deepening of water depths relative to the underlying supratidal facies at the top of step 4. The uppermost part (18 m) of step 8 consist of massive ledges (1–2 m) of wackestone that mark the subtidal carbonate environment also recorded during the final two isotope steps at Lawson Cove and Shingle Pass. Palmer (1979, 1984) located the extinction event marking the top of the Pterocephaliid biomere/Steptoean Stage in the House Range, and this horizon coincides with the beginning of the positive carbon isotope excursion marking the top of isotope step 8 (Fig. 4; Saltzman et al., 1995b).

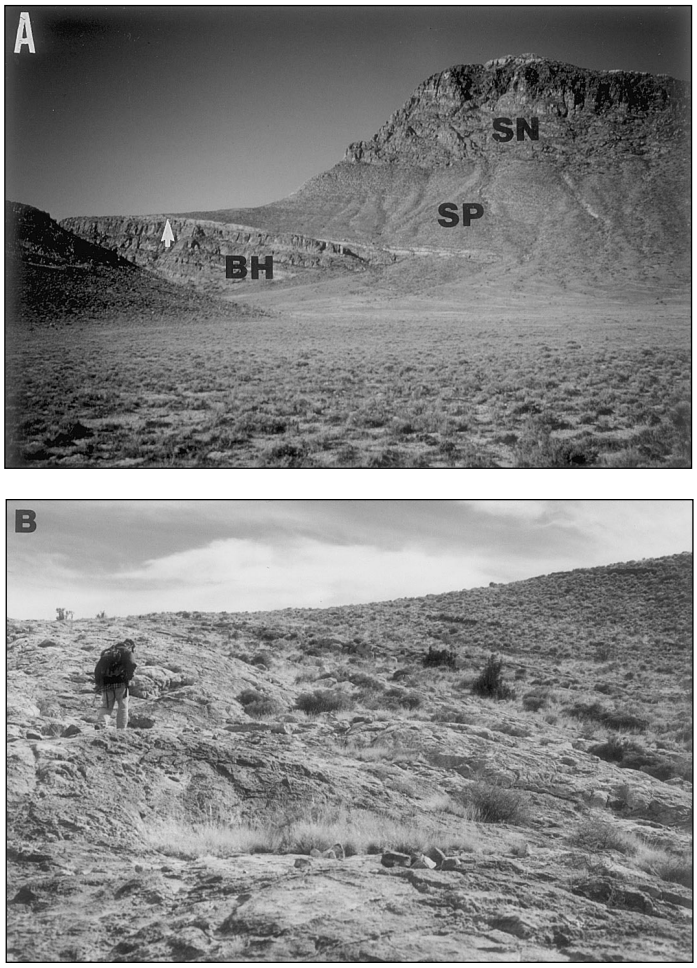

Figure 7. View of carbonate-siliciclastic transition (drowning surface) near the base of the Pterocephaliid biomere and beginning of the SPICE excursion. Lawson Cove section. (A) Section view showing recessive siliciclastics of the Steamboat Pass Shale (SP) between the Big Horse Limestone (BH) and Sneakover Limestone (SN). Arrow at top of Big Horse Limestone marks drowning surface. Area marked by arrow is shown in detail in bottom photograph. (B) Geologist for scale is at position of arrow in 7A.

DISCUSSION

Although variations in water depth and sediment type occurred frequently during the time of the SPICE excursion, only two such changes can be correlated among sections and seem to have regional significance. These changes occur near the beginning and near the peak of the SPICE excursion and correspond to abrupt carbonate-siliciclastic (or siliciclastic-carbonate) transitions. The transitions must reflect significant changes in sediment supply, subsidence, climate, or eustatic sea level that may ultimately provide clues to the nature of the global oceanographic events indicated by the carbon isotope excursion and mass extinction of trilobites.

Base of Pterocephaliid Biomere. At Shingle Pass, peritidal carbonates abruptly replace shaly intrashelf basin facies at a level 6 m above the base of isotope step 1 and mass extinction horizon. The shale-dominated environments characterized the area during the preceding *Cedaria* and *Crepicephalus* zones of the Marjumiid biomere (Member B of the Emigrant Springs Limestone of Kellogg, 1963) and their disappearance until well into isotope step 4 marks a major change in the pattern of sedimentation on the shelf. The House Range and Lawson Cove sections also record a major change in the pattern of sedimentation near the beginning of step 1. However, in contrast to the siliciclastic-carbonate transition observed at Shingle Pass, this change is an abrupt carbonate-siliciclastic transition (Fig. 7) that Palmer (1981a) termed Grand Cycle Top 10. This transition, which occurs 5 m below the beginning of step 1 in the House Range and between 1 and 6 m above it at Lawson Cove, marks the termination of a healthy shallow-water carbonate factory that characterized the preceding *Cedaria* and *Crepicephalus* zones in these areas (Big Horse Limestone; Hintze and Palmer, 1976). The overlying siliciclastics are the initial deposits of the deep ramp or intrashelf basin environments that characterize these areas well into step 7 at Lawson Cove, and step 4 in the House Range.

Osleger and Read (1993) interpreted the influx of siliciclastics and coincident drowning of the carbonate platform in the House Range to reflect a sequence boundary for which they find evidence in Texas and the Appalachian region. Their third-order sea-level curve shows this sequence boundary to correspond to a minimum in water depth on the platform, with the overlying shaly deposits (Candland Shale; Fig. 2) accumulating as sea level began to rise again. This sea-level curve is based in part on the paleobathymetric interpretation of the siliciclastic-dominated sediments as deeper water deposits relative to the underlying carbonates. To account for this counterintuitive movement of (cratonally derived) siliciclastics into the carbonate platform during rising sea level, several authors have invoked a model in which siliciclastics previously trapped in low-energy coastal basins are remobilized and carried onto the platform during storms (Mount and Rowland, 1981; Osleger and Montanez, 1996).

An alternative interpretation of the carbonate-siliciclastic transitions in the House Range and Lawson Cove sections is that falling sea level led to regression and an increase in the supply of siliciclastics. This may have decreased shallow water carbonate production and resulted in relative deepening on the shelf as subsidence outpaced siliciclastic sediment supply (Lochman-Balk, 1970; Thomas, 1993). In sequence stratigraphic terminology (Van Wagoner et al., 1987), the siliciclastics of the Candland Shale

would relate to a lowstand systems tract versus a transgressive systems tract in the scenario of Osleger and Read (1993).

Several additional observations bear on this debate over rising versus falling sea level near the beginning of the SPICE excursion. First, climatic shifts in cratonal source regions may play a significant role in controlling the supply of siliciclastics to the carbonate platform, independent of the position of eustatic sea level (Cowan and James, 1993; Osleger and Montanez, 1996). Second, local oceanographic conditions and topography on the shelf must play a significant role based on the fact that, as described above, the Shingle Pass section does not record a coeval carbonate-siliciclastic transition. Rather, an abrupt siliciclastic-carbonate transition occurs above the base of the isotope step 1. A shift to carbonate-dominated sedimentation is also observed in the southern Canadian Rockies near the base of the Pterocephaliid biomere/Steptoean Stage (the midcycle boundary of the Sullivan-Lyell Grand Cycle; Aitken, 1966, 1978). The precise timing of these siliciclastic-carbonate transitions relative to the opposite transitions in the House Range and Lawson Cove sections is not known, but this would seem to caution against the use of the occurrence of siliciclastics to infer the position of third-order sea level (Chow and James, 1987; Cowan and James, 1993; Osleger and Read, 1996) without good constraints on the paleogeography of the region.

Paleogeography is partly determined by the role of local tectonic factors, and Rees (1986) inferred a northeast-southwest–trending normal fault through the carbonate platform in the eastern Great Basin, based on abrupt lateral facies changes in Middle Cambrian strata. This fault runs within 50 km of the House Range, Lawson Cove, and Shingle Pass sections. Reactivation of the fault, which created a trough known as the House Range Embayment during Middle Cambrian time (Rees, 1986), may have diverted the supply of siliciclastics from the Shingle Pass area to the House Range and Lawson Cove sections. Faulting may also have resulted in differential subsidence on the platform. For example, during isotope step 2 higher rates of subsidence and/or sediment supply produced a greatly thickened siliciclastic sequence in the House Range relative to Lawson Cove, consistent with a position closer to the axis of the House Range Embayment.

A notable siliciclastic-carbonate transition in the House Range section during isotope step 4 may also reflect a diversion of clastic sediment supply or, in addition, may reflect progradation of the shoal-water complex into the axis of the House Range Embayment. The shallow-water carbonate factory was not similarly reestablished at the Lawson Cove section, which is extremely condensed during isotope step 4 and records 4 m of strata versus the 70 m that accumulated in the House Range. A decrease in the supply of fine clastics from the craton or differential subsidence seems likely to have played a role in condensation.

Sauk II–Sauk III Boundary. Carbonate facies in the House Range and Shingle Pass sections shallow to supratidal depths toward the end of isotope step 4, near the peak of the SPICE excursion. The shallow-water carbonates are then drowned by an influx of siliciclastics. This regionally significant carbonate-siliciclastic transition, referred to as the Sauk II–Sauk III event (Palmer, 1981b) presents similar interpretational difficulties that have been discussed above. However, the Sauk II–Sauk III carbonate-siliciclastic transition appears different in that it is tied to a major sedimentary hiatus in cratonal sections of North America. This hiatus was proposed by Palmer (1981b) to explain missing trilobite zones in cratonal regions of North America. The carbon isotope stratigraphy through the Pterocephaliid biomere/Steptoean Stage in cratonal sections in Wyoming do not record the SPICE excursion (Fig. 8), providing independent confirmation of this major sedimentary hiatus. Exposure and erosion or nondeposition occurred during the time represented by isotope steps 2–7 in Wyoming.

Palmer (1981b) proposed that the Sauk II–Sauk III hiatus sections may reflect a eustatic fall in sea level. This notion has been embraced by Cambrian stratigraphers across North America (James et al., 1989; Aitken, 1993), and Osleger and Read (1993) considered the Sauk II–Sauk III event to be an unequivocal Type 1 sequence boundary. However, physical evidence for a synchronous drop in sea level is lacking (e.g., incised valleys of Christie-Blick, 1995). In fact, a recent study of the stratigraphy of the Pterocephaliid biomere/Steptoean Stage in the nearshore sandstones of the Upper Mississippi Valley (Runkel et al., 1996) did not reveal convincing evidence for subaerial exposure. It is thus possible that the remarkably widespread change to a siliciclastic depositional regime near the top of isotope step 4 in much of the Great Basin and the condensation or hiatus recognized in more cratonal sections (Chow and James, 1987) are not related to sea-level drop.

One alternative is that this transition marks a decrease in the carbonate saturation state of the oceans. Dissolution and/or reduced rates of carbonate accumulation in such an ocean could contribute to the dominantly siliciclastic succession recorded by the Corset Spring Shale (Fig. 2) in the Great Basin (or the Dry Creek Shale in Wyoming). Such a change in ocean chemistry might be related to oxidation of previously stored organic matter and is consistent with the observation that the SPICE excursion began to turn around from peak $\delta^{13}C$ values near the top of isotope step 4. The oxidation of significant amounts of organic matter is one way to lower $\delta^{13}C$ ratios. Other ways include decreasing the $\delta^{13}C$ of the riverine carbon flux or decreasing the burial ratio of organic carbon to carbonate carbon. Subsequent termination of siliciclastic deposition during isotope steps 7 and 8, as blanket subtidal carbonates spread across the study area and throughout the Great Basin (Brady and Rowell, 1976), must also reflect a significant change in eustatic sea level, subsidence, sediment supply, or ocean chemistry.

The observed changes in $\delta^{13}C$, sedimentation patterns, and the marine fauna during the SPICE excursion record major paleoceanographic events. In order to better understand the significance of these events, it is instructive to consider both the regional expression in other parts of Laurentia and the uniqueness of this event within the intensively studied Upper Cambrian rocks of the Great Basin.

Regional Significance

The SPICE excursion is reflected in a remarkably wide range of carbonate lithologies, including oolitic grainstones, burrowed wackestones, thrombolitic boundstones, laminated micrites and flat-pebble conglomerates, which represent various positions and water depths across the shelf and will have undergone varying degrees of diagenesis. Thus, carbonate strata deposited during any portion of Pterocephaliid biomere/Steptoean Stage and its time equivalents will record that step of the SPICE excursion, and the chronostratigraphic framework outlined in Figure 4 may be applied elsewhere (Fig. 9). Such an approach may help answer questions concerning the regional significance of the stratal surfaces near the base of isotope step 1 (base of the Pterocephaliid biomere) and the top of isotope step 4 (Sauk II–Sauk III boundary). In fact, major changes in sedimentation patterns approximately correlative with these time intervals have been observed in the southern Appalachians (Osleger and Read, 1993), western Newfoundland (Chow and James, 1987), and the southern Canadian Rocky Mountains (Aitken, 1981).

In the Conasauga basin of the Tennessee-Virginia Appalachians, the base of the Pterocephaliid biomere/Steptoean Stage (*Aphelaspis* zone time) is marked by the widespread expansion of shales of the upper Nolichucky Formation from an intrashelf basin onto the surrounding peritidal platform (Markello and Read, 1982; Osleger and Read, 1993). The Nolichucky shales grade into ribbon rocks and cryptalgal laminites of the Maynardville Formation, above which Osleger and Read (1993) recognized the Sauk II–Sauk III event as a very thin interval in the basal Copper Ridge (Conococheague) Formation. This thin interval is characterized by abundant quartz sand, thin peritidal cycles, and subtle erosional truncation of

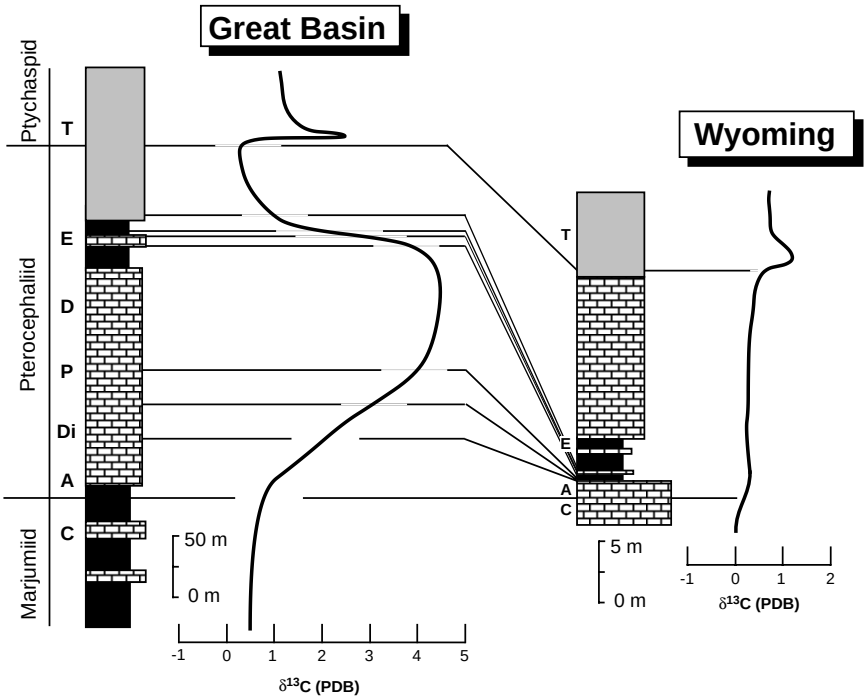

Figure 8. Schematic comparison of $\delta^{13}C$ curves from the Great Basin and cratonal sections of Wyoming (Saltzman and Lohmann, unpublished data). Isotope step boundaries from Figure 4 are shown as solid lines. As indicated with capital letters, all five fossil zones of the Pterocephaliid biomere (see Fig. 3 for zone names) are known from the Great Basin and only two are known from Wyoming. Lithologic patterns are from Figures 3 and 4.

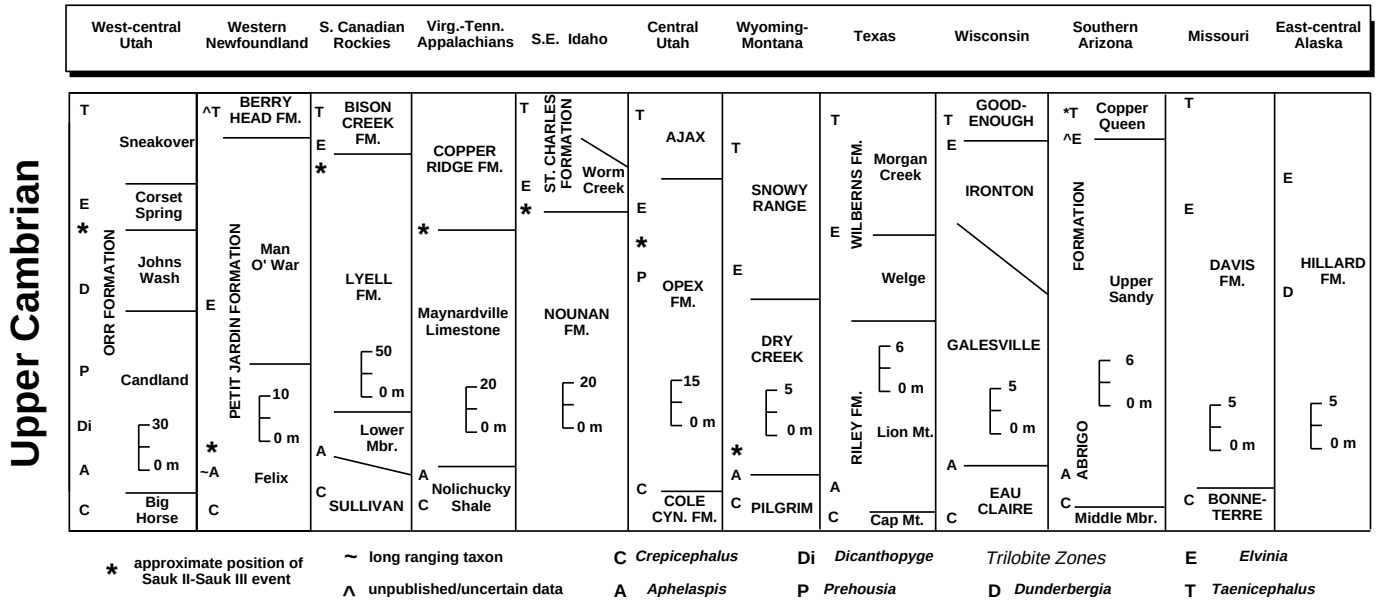

Figure 9. Correlation chart for the Upper Cambrian of Laurentia showing approximate positions of trilobite collections relative to formation boundaries. We have tried to place the position of the Sauk II–Sauk III event relative to formational boundaries based either on claims made in published reports or simply by the report of an abrupt carbonate-siliciclastic transition. The following sources were used to compile this chart: west-central Utah (Palmer, 1965b; Hintze and Palmer, 1976); western Newfoundland (Chow and James, 1987; Westrop, 1992); southern Canadian Rockies (Aitken, 1978, 1981; Westrop, 1989); Virginia-Tennessee Appalachians (Markello and Read, 1982; Osleger and Read, 1993); southeast Idaho (Williams and Maxey, 1941; Williams, 1948; Palmer, 1971); central Utah (Koepnick, 1976; Palmer, 1971); Wyoming-Montana (Lochman-Balk, 1970, 1971; Saltzman et al., 1995); Texas (Wilson, 1949; Lochman-Balk, 1971); Wisconsin (Raasch and Unfer, 1964; Lochman-Balk, 1971); southern Arizona (Gilluly, 1956; Lochman-Balk, 1971); Missouri (Lochman-Balk, 1971); east-central Alaska (Palmer, 1968).

some cycle tops, although unequivocal evidence for a major exposure surface is lacking.

In western Newfoundland, Westrop (1992) was able to constrain the base of the Pterocephaliid biomere/Steptoean Stage to within 5 m of section. Although no major facies changes are noted, several beds of quartzose sandstone occur about 5 m above the base of the Pterocephaliid biomere/Steptoean Stage, and Chow and James (1987) considered strata enclosing these beds to represent a condensed interval that formed as a result of a slowing of sea-level rise during the time of the Sauk II–Sauk III event. Cowan and James (1993) suggested that this quartz rich horizon reflects marine reworking of eolian sand during a low-stand of sea level.

In the Southern Canadian Rockies, the base of the Pterocephaliid biomere/Steptoean Stage occurs near the Sullivan-Lyell formation boundary, which marks a major siliciclastic-carbonate transition (Aitken, 1978). Aitken (1981) has also noted a pronounced lithologic change near the top of the carbonates of the Lyell Formation, which he suggested may represent the Sauk II–Sauk III event. Such hypotheses could be tested using the carbon isotope stratigraphy outlined in this paper.

Stratigraphic Context

Prior to the abrupt facies changes near the base of the Pterocephaliid biomere/Steptoean Stage (Grand Cycle Top 10, Palmer, 1981a), ~400–500 m of shelf sediments were deposited without any distinctive changes in the pattern of sedimentation (Drewes and Palmer, 1957; Kellogg, 1963; Hintze and Robison, 1975). This stability followed a regional facies change near the Middle–Late Cambrian boundary known as Grand Cycle Top 9 (Palmer, 1981a). Grand Cycle Top 9 in the Great Basin is marked by widespread siliciclastic influx (Fish Springs Member of the Trippe Limestone; Hintze and Robison, 1975) and locally large-scale submarine gravity flows (Member A of Emigrant Springs Limestone; Kellogg, 1963; Palmer, 1971). These distinctive facies changes seem to reflect a regional tectonic, climatic or sea level event, although they are not associated with extinctions and the carbon isotope record is only poorly known (Brasier, 1993). The next older notable carbonate-siliciclastic transition occurred at the base of the Middle Cambrian *Ptychagnostus gibbus* biochron and has been recognized on at least five continents (Robison, 1976, 1984). This global event, which also marks the initial development of the House Range embayment in the Great Basin (Rees, 1986), is associated with only minor faunal turnover and little change in the carbon isotope ratios of carbonate rocks (Saltzman and Lohmann, unpublished).

A significant interval of extinctions at the top of isotope step 8 (Pterocephaliid-Ptychaspid biomere boundary) is associated with an ~2‰ change in carbon isotope ratios in some sections (Saltzman et al., 1995b), but this excursion is brief and not associated with major facies changes in the eastern Great Basin. Above this excursion, another ~500 m of platform carbonates are deposited (Notch Peak Formation) before an apparent lowering of sea level near the base of the Ordovician (Miller, 1992). This event is associated with significant extinctions and a minor change in carbon isotope ratios in Australia (Ripperdan et al., 1992). Thus, it is the exceptional magnitude and duration of the changes in carbon cycling implied by the SPICE excursion and their association with the Sauk II–Sauk III hiatus on the craton that thus far are unique within Cambrian time.

Global Significance

It would seem that the most likely scenario to explain the initiation of the SPICE excursion involves the destabilization of a previously stratified Cambrian ocean (Taylor, 1977). This might have occurred as a result of a combination of climatic, eustatic and tectonic factors and would potentially provide a mechanism for simultaneously perturbing the carbon cycle and the marine biota (Wilde and Berry, 1984; Wilde et al., 1990). The absolute age and duration of the SPICE excursion are unknown. It is younger than 500 Ma, based on U-Pb ages on volcanic rocks in Tasmania (Perkins and Walshe, 1993), and older than the 490–495 Ma age of beginning of Ordovician time (Jones, 1994; Tucker and McKerrow, 1995). Given that there are 10–15 trilobite zones in Late Cambrian time in most regions (Peng, 1992), and that the SPICE excursion lasted for about five trilobite zones (Palmer, 1965b), its duration is tentatively estimated to be 2–4 Ma. This is longer than the mean residence time of carbon in the oceans (~350 ka) and suggests that the excursion resulted from a change in the steady-state partitioning of oceanic carbon between the sedimentary organic carbon and carbonate carbon reservoirs (Shackleton, 1987; Kump, 1991).

Specifically, an increase in the burial ratio of organic carbon relative to carbonate carbon would remove light (^{12}C) carbon from the global oceans as organic-rich sediments, thereby causing the positive $\delta^{13}C$ increase we observe. Although we do not see convincing evidence for this black shale deposition in the carbonate platform sections we have studied in the Great Basin, Thickpenny and Leggett (1987) single out the Late Cambrian Epoch as a time of widespread organic-rich deposition in Europe. If organic-rich deposition is the cause of the SPICE excursion, then an increase in primary productivity or decrease in the oxygenation state of ocean basins may have occurred (Arthur et al., 1987). Van Cappellen and Ingall (1996) showed that burial of the limiting nutrient, phosphorus, is less efficient when bottom waters are low in oxygen, and this idea could provide a mechanism to sustain the millions of years of high productivity and low oxygen conditions that seem to be necessary to account for the SPICE excursion. Alternatively, or in addition to organic matter burial, a sink for the light (^{12}C) carbon might be in the form of methane hydrates (Dickens, et al., 1995). These CH_4 hydrates are extremely enriched in ^{12}C ($\delta^{13}C = -60‰$) relative to carbonates or organic matter ($\delta^{13}C = -25‰$) and thus their formation would have a significant effect on oceanic $\delta^{13}C$. Global cooling is one way to increase the amount of methane hydrates stored in the oceans.

Using modern values for carbon fluxes and the isotopic fractionation due to photosynthesis (Arthur et al., 1987), we calculate an excess rate of burial of organic carbon (ignoring methane hydrates) of about 10^{19} g C Ma^{-1} over the duration of the SPICE excursion. This amount (10^{19} g) is an order of magnitude more than the mass of carbon in the modern atmosphere ($\approx 10^{18}$ g). If rates of CO_2 release to the ocean-atmosphere system due to outgassing and weathering did not change, this excess burial of organic carbon could have resulted in a significant reduction in atmospheric CO_2 and, as a result, may have cooled the Earth through a reversal of the greenhouse effect (Arthur et al., 1988). The question remains whether the Sauk II–Sauk III hiatus is related to a change in sea level or ocean chemistry that resulted ultimately from an episode of global cooling.

The rapid return of $\delta^{13}C$ values to preexcursion levels may reflect CO_2-induced global cooling in that decreased temperatures should have led to an increase in the oxygen concentration of deep water. Acting together with a fall in sea level, this increased oxygenation state may have led to oxidation of previously stored organic matter. There is, however, no evidence for Late Cambrian ice-sheets, and perhaps changes in the amount of water stored in continental reservoirs related to global cooling are necessary to produce a significant fall in sea level (e.g., Jacobs and Sahagian, 1993). The prediction of CO_2-induced climatic cooling and a carbon cycle control on sea level during the SPICE excursion might be testable with more reliable oxygen isotope data (e.g., Vincent and Berger, 1985; Brenchley et al., 1994). In addition, measurements of the difference between the isotopic compositions of carbonate and coeval organic matter may provide constraints on the predicted decrease in the level of carbon dioxide in the atmosphere (Arthur et al., 1988).

CONCLUSIONS

We have shown that $\delta^{13}C$ ratios of ancient seawater are recorded in a wide range of carbonate lithologies from the Great Basin. Arbitrary 1‰ divisions of a nearly 5‰ secular trend in $\delta^{13}C$ during the Pterocephaliid biomere/Steptoean Stage define eight isotope steps, which are useful in correlation of two important stratal surfaces. The first of these surfaces occurs near the beginning of isotope step 1 and the second near the end of isotope step 4. These surfaces correspond to changes between siliciclastic and carbonate dominated sedimentation on the shelf due to a combination of tectonic, climatic, and eustatic influences. The observed changes in $\delta^{13}C$, sedimentation, and the marine fauna during the SPICE excursion provide remarkably detailed records of a major paleoceanographic event.

The establishment of carbon isotope profiles during the Pterocephaliid biomere/Steptoean Stage elsewhere should improve chronostratigraphic correlation during Late Cambrian time and ultimately allow for an unprecedented view of sedimentation patterns and biological evolution on an ancient carbonate platform. It should also result in more precise correlation between the Great Basin and regions such as the Appalachians or the Canadian Rockies, where the sedimentology is also well known but the resolution of the fossil record is currently an order of magnitude worse than in the Great Basin region. Such a level of resolution will be necessary to test hypotheses regarding the observed paleoceanographic events during the time of the SPICE excursion.

On a global scale, our results confirm that large perturbations in the carbon isotope ratios of common carbonate rocks may be used as precise measures of time that transcend biogeographic barriers (Kaufman et al., 1991; Brasier et al., 1994; Kaufman and Knoll, 1995). The results will be a major step toward the ultimate goal of a high-resolution record of parallel changes in isotopes, sea level, and marine fauna that may be critical to understand linkages among the carbon cycle, sea level, and the evolution of the biosphere.

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