THE LATE CAMBRIAN SPICE ($\delta^{13}C$) EVENT AND THE SAUK II–SAUK III REGRESSION: NEW EVIDENCE FROM LAURENTIAN BASINS IN UTAH, IOWA, AND NEWFOUNDLAND

MATTHEW R. SALTZMAN,[1] CLINTON A. COWAN,[2] ANTHONY C. RUNKEL,[3] BRUCE RUNNEGAR,[4] MICHAEL C. STEWART,[2*] AND ALLISON R. PALMER[5]

[1] *Department of Geological Sciences, Ohio State University, Columbus, Ohio 43210, U.S.A.*
email: saltzman.11@osu.edu
[2] *Department of Geology, Carleton College, Northfield, Minnesota 55057, U.S.A.*
[3] *Minnesota Geological Survey, St Paul, Minnesota 55114-1057, U.S.A.*
[4] *Department of Earth and Space Sciences and Astrobiology Institute, University of California, Los Angeles, California 90095, U.S.A.*
[5] *Institute for Cambrian Studies, 445 North Cedarbrook Road, Boulder, Colorado 80304-0417, U.S.A.*

ABSTRACT: Carbon isotope data from Upper Cambrian sections in three Laurentian basins in northern Utah, central Iowa, and western Newfoundland record a large positive $\delta^{13}C$ excursion (SPICE event) of up to + 5‰. Peak $\delta^{13}C$ ratios are well dated by trilobite collections to the middle of the Steptoean Stage (*Dunderbergia* Zone) and occur during maximum regression associated with formation of the Sauk II–Sauk III subsequence boundary on the North American craton. Maximum regression was marked by an influx of quartz sand into carbonate-platform settings in all three widely separated basins. In northern Utah, this quartz sand formed a thick sequence known as the Worm Creek Quartzite, which marks a conspicuous interruption of carbonate deposition during the Middle to Late Cambrian in the region. In western Newfoundland, the thickness of the quartz sand unit is much reduced but still marks a brief shutdown of the carbonate factory that is unique to the Cambrian shelf succession of the area. In the central Iowa area of the cratonic interior, an upward-shallowing carbonate succession culminates in cross-stratified trilobite grainstones at the peak of the SPICE in *Dunderbergia* Zone time, and the lowest point on the relative-sea-level curve is associated with the occurrence of coarse quartz sand derived from the encroaching shoreface.

Although it is difficult to determine precisely the departure from baseline $\delta^{13}C$ that marks the beginning of the SPICE excursion in the stratigraphic successions analyzed, our results are consistent with a rise and subsequent fall in $\delta^{13}C$ tracking a major regressive–transgressive event recorded across northern Laurentia. The correlation of a major $\delta^{13}C$ excursion with regression is similar to that described for the Late Ordovician, for which the pattern has been attributed to either increased carbonate relative to terrigenous weathering rates as ice sheets covered up organic-matter-containing silicates at high latitudes or high productivity and organic-carbon burial driven by oceanic overturn. The lack of known Steptoean-age ice sheets that could have affected the ratio of carbonate to silicate weathering rates suggests that organic-carbon burial was the likely cause of the SPICE event. We suggest that increased weathering and erosion rates during relative sea-level fall (Sauk II–III) increased the burial fraction of organic carbon in an expanded region of fine-grained siliciclastic deposits in shelf and upper slope environments during the Steptoean.

INTRODUCTION

The Steptoean positive carbon isotope excursion (SPICE) is characterized by a positive $\delta^{13}C$ shift of ~+ 4‰ measured in Upper Cambrian carbonate rocks (Steptoean Stage) in North America and time-equivalent strata in China, Kazakhstan, and Australia (Fig. 1; Saltzman et al. 2000). The SPICE is recorded across > 200 m of shallow-water limestones in sections in Nevada and likely represents a perturbation in the carbon cycle that lasted for several million years or less on the basis of best estimates of Late Cambrian accumulation rates (Osleger and Read 1993) and absolute age constraints (e.g., Davidek et al. 1998). Positive carbon isotope excursions of similar magnitude and duration are known to have occurred on average about once per each successive Paleozoic period (Late Ordovician, Late Silurian, Frasnian–Famennian, Early Mississippian, and early Late Permian events; Gruszczynski et al. 1989; Brenchley et al. 1994; Joachimski et al. 2002; Saltzman 2001, 2002). These transient $\delta^{13}C$ peaks are traditionally interpreted to reflect increased burial of isotopically light (^{12}C-enriched) organic carbon, which has been linked in different paleoceanographic models to increased productivity, enhanced preservation under anoxic conditions, or high sedimentation rates (e.g., Arthur et al. 1987; Derry et al. 1992; Schrag et al. 2002). Adsorption of organic carbon onto clay-mineral surfaces may also be an important variable in the burial and preservation of organic matter on geological timescales (Kennedy et al. 2002). For $\delta^{13}C$ peaks that persist for several million years or more, the development of anoxic conditions in regions of high rates of organic-matter burial appears to play a key role by increasing the carbon-to-phosphorus ratio of the buried organic matter (Van Cappellen and Ingall 1994).

An alternative model relates positive excursions in $\delta^{13}C$ to transient increases in the weathering of isotopically heavy carbonate-platform deposits relative to fine-grained terrigenous rocks that contain abundant organic matter (Kump et al. 1999). This change in the weathering ratio can occur as carbonates are preferentially exposed at low latitudes during a glacio-eustatic fall, while at the same time expanding ice sheets may cover organic-matter-containing terrigenous rocks in high latitudes. The $\delta^{13}C$ of the riverine input may increase in such a scenario, although the counterbalancing effects of, for example, exposing and eroding previously submerged distal portions of siliciclastic shelves at high latitudes are difficult to evaluate.

To the extent that marine inundation of the continents influences weathering rates and the input and output ratios of oxidized and reduced carbon, changes in sea level must have an important influence on seawater $\delta^{13}C$. In comparing peak $\delta^{13}C$ values with the position of sea level, two contrasting relationships have been documented in the literature. For example, $\delta^{13}C$ maxima in the Late Ordovician and Cenomanian–Turonian boundary intervals are well known to coincide with major lowstand and highstand positions of sea level, respectively (Arthur et al. 1987; Finney et al. 1999). Rising sea level apparently caused the Cenomanian–Turonian event through circulation-driven increases in production and preservation of organic matter, a model also proposed for the Frasnian–Famennian boundary $\delta^{13}C$ excursion (Joachimski and Buggisch 1993; Joachimski et al. 2002). On the other hand, recognition of a glacio-eustatic drop during the Late Ordovician event (Brenchley et al. 1994) may indicate that changing weathering fluxes increased riverine $\delta^{13}C$ values (Kump et al. 1999) or led to overall higher sedimentation rates that enhanced preservation of organic matter. Higher weathering rates of silicate- or phosphate-rich sedimentary rocks during regression may potentially increase nutrient delivery to shelf regions and enhance primary production and carbon burial. Increased phosphate deliv-

* Present address: Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts 02139, U.S.A.

JOURNAL OF SEDIMENTARY RESEARCH, VOL. 74, NO. 3, MAY, 2004, P. 366–377
 1527-1404/04/074-366/$03.00

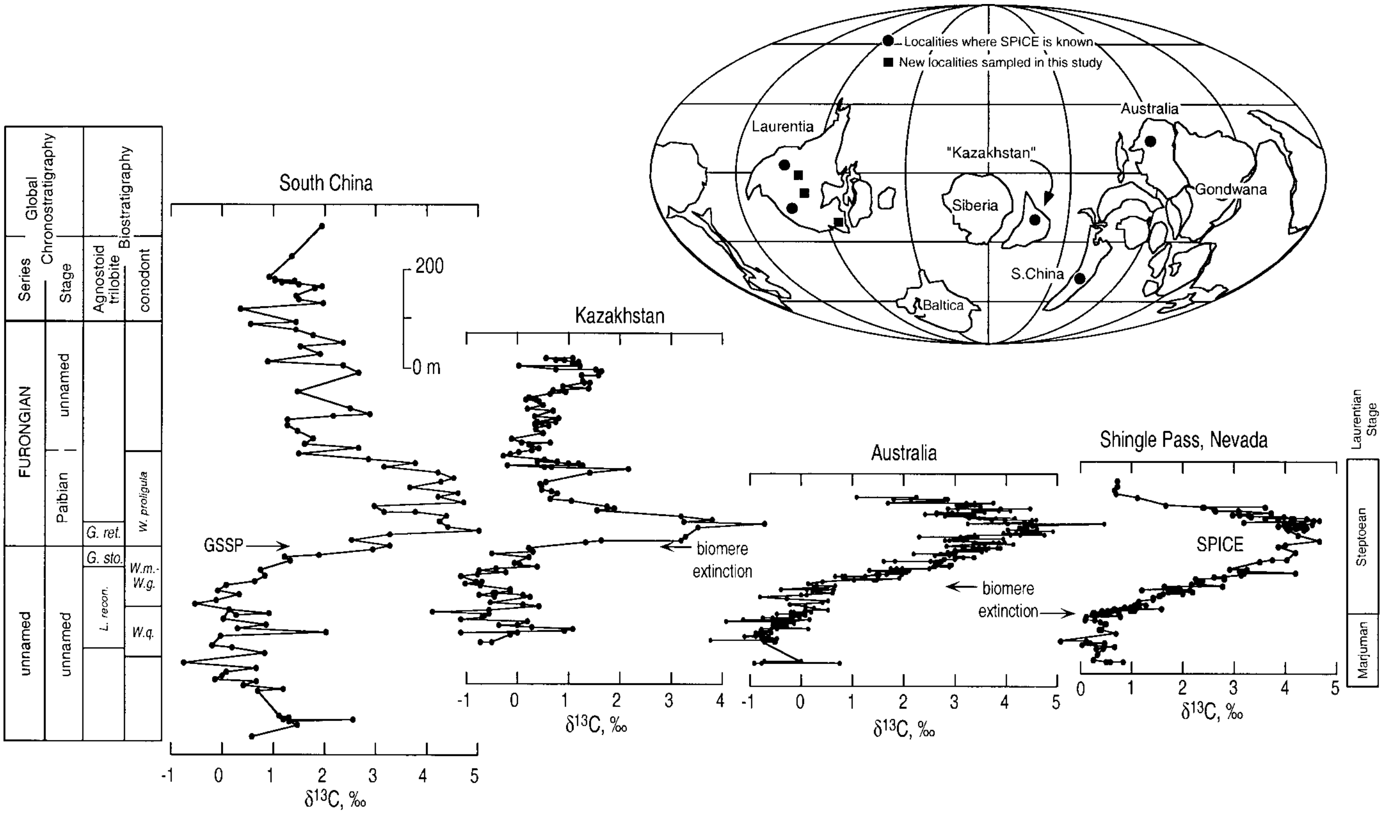

FIG. 1.—Cambrian record of the SPICE excursion and global chronostratigraphic framework (Peng et al. 2001). Carbon isotope profiles based on data in Saltzman et al. (2000a), with paleogeographic map modified to include localities sampled in this investigation, as well as the southern Appalachian (Laurentian) locality in which the SPICE is known (Glumac and Walker 1998).

ery from the erosion of continental rocks has also been argued for highstands of sea level and associated $\delta^{13}C$ peaks in the Cretaceous due to warm, humid conditions and an accelerated water cycle (Föllmi et al. 1994).

The peak of the SPICE $\delta^{13}C$ event in parts of Laurentia (Glumac and Walker 1998; Saltzman et al. 1998) appears to coincide with a sea-level lowstand known as the Sauk II–III boundary (Palmer 1981). The Sauk II–III sequence boundary has been recognized at a number of sites across Laurentia by use of both faunal and lithic criteria that provide evidence for a period of erosion or nondeposition within the long-term Sauk sequence of Sloss (1963). Osleger and Read (1993) considered the Sauk II–III boundary to be the only unequivocal Type 1 sequence boundary in the Upper Cambrian. However, because the precise position of this sequence boundary in thick and conformable, mixed carbonate–siliciclastic successions is not always clear, a causal connection with the SPICE event remains uncertain. For example, Steptoean sections in Utah and Nevada that record the SPICE show striking local variations in the nature and thickness of carbonate facies associations, which complicate the interpretation of relative sea-level trends in relation to $\delta^{13}C$ trends (Saltzman et al. 1998). In the section at Thorn Hill in eastern Tennessee examined by Glumac and Walker (1998), the influx of quartz sand into tidal-flat carbonates containing multiple cycle-capping exposure surfaces marks the Sauk II–III boundary as a zone (correlative conformity interval) within which the $\delta^{13}C$ peak occurs.

The purpose of this paper is to further evaluate the evidence for the temporal correspondence of a major regressive event (Sauk II–III boundary) with the SPICE by focusing on new data from three widely separated Laurentian basins in Utah, Iowa, and Newfoundland. Faunal control is good to excellent in all three of the sampled sections, with all of the major Steptoean trilobite zones recognized, and in all three basins we have found a striking coincidence between the position of the Sauk II–III boundary and the peak of the SPICE event. With no evidence for major ice sheets that changed the weathering ratio of carbonate to organic-matter-bearing shales, it seems likely that a net increase in organic-carbon burial in expanded areas of fine-grained siliciclastic deposits in submerged outer shelf environments caused the SPICE.

GEOLOGIC BACKGROUND

Laurentian Paleogeography

During the Late Cambrian, Laurentia was in an equatorial position with the Cordilleran and Appalachian passive margins oriented in east–west directions at positions $\sim$15° N and S in latitude, respectively (Fig. 2). The thick ($\leq$ 2 km), wedge-shaped passive-margin prisms were characterized by broad regions of dominantly shoal-water carbonate facies (middle carbonate belt) that passed oceanward into a deeper-water siliciclastic facies association (outer detrital belt) and landward into siliciclastic deposits of the cratonic interior (inner detrital belt). With global sea level near a Phanerozoic maximum as a result of both the greenhouse climate (Berner 1990) and the effects of continental breakup on ocean-ridge volumes (Bond et al. 1984), the paleo-shoreface was located far inland in the area of present-day Wisconsin, and marine waters inundated previously exposed cratonic embayments. The three studied sections (Fig. 2) represent the Cordilleran (Utah) and Appalachian (Newfoundland) passive margins and the intervening interior craton of the Upper Mississippi Valley (Iowa), thus providing a cross section of depositional environments spanning a line $\sim$5000 km in length across the Laurentian craton.

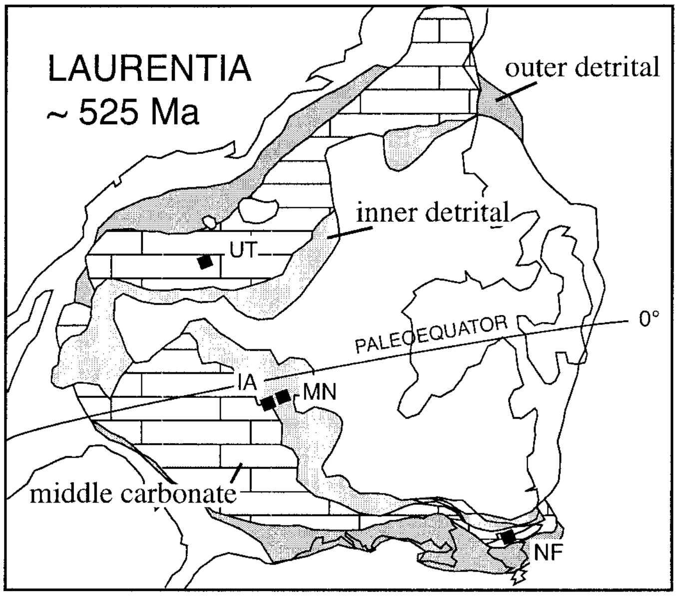

FIG. 2.—Facies map of Laurentia showing localities investigated in this study and their position with respect to Cambrian facies belts (Lochman-Balk 1970; Palmer 1971). UT = Smithfield Canyon, Utah; IA = Rhinehart core, Iowa; MN = Minnesota outcrop belt (see Runkel et al. 1998); NF = Felix Cove, Newfoundland.

Laurentian Biostratigraphy and the Sauk II–III Sequence Boundary

The Steptoean Stage (equivalent to the Pterocephaliid biomere of Palmer 1998) consists of five trilobite zones that were defined in the shelf and slope sequences of the central Great Basin (Palmer 1965), but outside of this region a tripartite scheme that includes the *Aphelaspis, Dunderbergia,* and *Elvinia* zones has been used (Fig. 3). Steptoean successions in several cratonal settings do not contain fossils representative of the *Dunderbergia* Zone, which is presumed to be missing at a hiatal surface formed as a result of the Sauk II–III regression (Fig. 3; Palmer 1981). The existence of a significant sedimentary hiatus during the Steptoean is further substantiated by carbon isotope stratigraphy, which shows that the SPICE excursion is missing in northwestern Wyoming (Saltzman et al. 1998) and northern Vermont (Glumac and Spivak-Birndorf 2002). Although little direct evidence for subaerial exposure and erosion exists at the Sauk II–III sequence boundary in the Upper Mississippi Valley, Runkel et al. (1998) used outcrop and subsurface biostratigraphically constrained facies relationships across six adjacent Midwestern states to estimate that the shoreline regressed about 300 km from western Wisconsin to central Iowa by late *Dunderbergia* Zone time. This corresponds to a mid-Steptoean sea-level fall of at most a few tens of meters, on the basis of an inferred shelf gradient of approximately 0.1 m/km (Runkel et al. 1998).

METHODS AND RESULTS

Homogeneous micrites were the preferred component in the drilling of polished slabs for geochemical analysis, although many of the samples (~1 mg of powder) isolated had a coarser-grained component that was recrystallized to various degrees. This reflects the wide range of carbonate lithologies that were sampled to generate high-resolution curves, including oolitic packstones, skeletal wackestones, thrombolitic boundstones, and flat-pebble conglomerates. The reliability of this sampling strategy is based on the good reproducibility of $\delta^{13}C$ trends generated with micrites for the same time periods, using independent biostratigraphic correlation, and from different depositional environments on separate cratons (e.g., Ripperdan et al. 1992; Saltzman et al. 1998; Saltzman et al. 2000; Kump et al. 1999; Joachimski et al. 2002). In addition, the methods employed here are supported by the similarity of features between micrite-based and brachiopod-based curves (which are often considered the most reliable in $\delta^{13}C$ stratigraphy) for periods such as the Silurian (e.g., Azmy et al. 1998; Saltzman 2001), and the Mississippian (Mii et al. 1999; Saltzman 2002). The results of quantitative modeling studies further show that $\delta^{13}C$ values are likely to be rock-buffered over a wide range of diagenetic settings commonly encountered in ancient carbonate successions, whereas $\delta^{18}O$ is easily reset (Lohmann 1988; Banner and Hanson 1990; Gao and Land 1991).

Despite the potential for recovery of primary values, diagenetic alteration of $\delta^{13}C$ is always a concern in the interpretation of matrix micrite data, particularly where shallowing-upward carbonate cycles capped by subaerial exposure surfaces are present and incorporation of soil-derived CO_2 is likely (e.g., Algeo et al. 1992; Railsback et al. 2003). In addition, it is important to target carbonates formed in relatively open-marine environments representing maximal exchange rates with the open ocean rather than the interior carbonate platforms, in which $\delta^{13}C$ may vary in proportion to the local residence time of seawater and the rate of organic-matter remineralization (Holmden et al. 1998; Immenhauser et al. 2002).

To minimize the degree of overprinting of the global seawater $\delta^{13}C$ record, the sections sampled in this study represent dominantly subtidal, open-ramp successions that lack obvious or frequent exposure features. This selectivity was not possible in every instance, and thus we have in part relied on the extent to which the sections display relatively steady $\delta^{13}C$ trends that can be correlated between widely separated sections. Carbonate powders in this study were roasted under vacuum at 380° C for one hour to remove volatile contaminants and reacted with 100% phosphoric acid at 75° C in an online carbonate preparation line (Carbo-Kiel; single-sample acid bath) connected to a Finnigan Mat 252 or 251 mass spectrom-

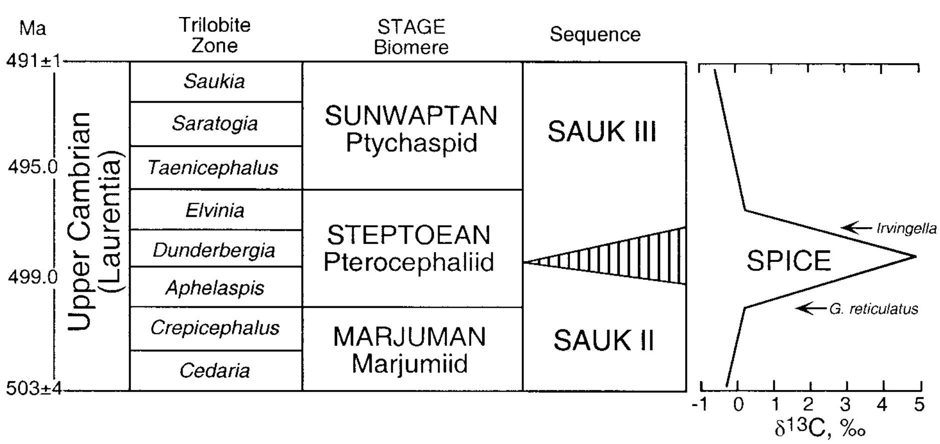

FIG. 3.—Upper Cambrian stratigraphic framework for Laurentia with trilobite zones modified from Palmer (1965), and stages and biomeres taken from Palmer (1998). Sauk sequence subdivisions is after Palmer (1981). Cosmopolitan trilobites *Glyptagnostus reticulatus* and *Irvingella* are used for correlation between Laurentia and other continents. Duration of the SPICE excursion (Saltzman et al. 1998) is obtained by interpolation between two U–Pb zircon ages (Perkins and Walshe 1993; Davidek et al. 1998).

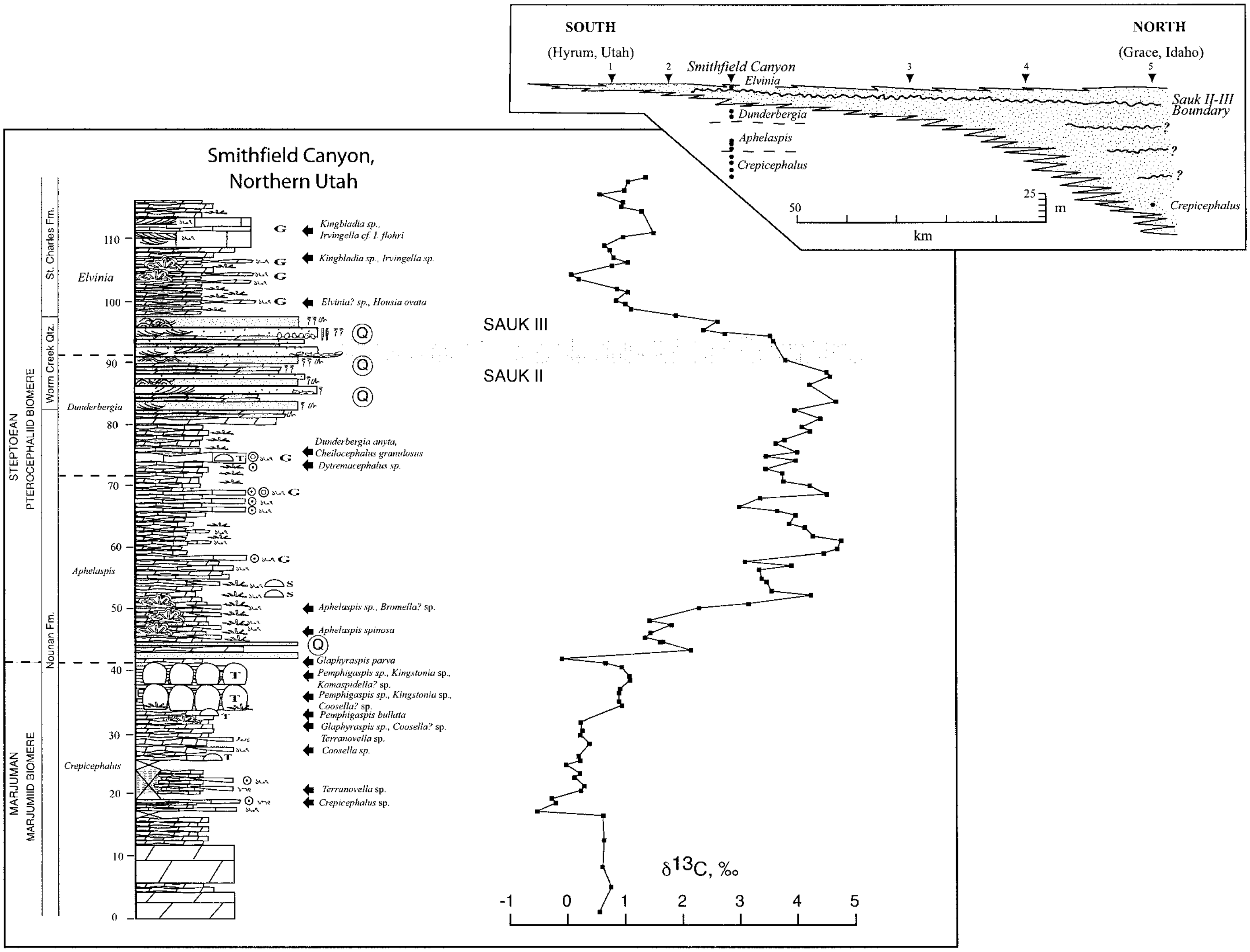

FIG. 4.—Carbon isotope stratigraphy and trilobite collections at Smithfield Canyon section, northern Utah. See lithologic legend in Figure 7. The Worm Creek Quartzite is sandwiched by subtidal carbonate shelf strata with subordinate quartz sandstone beds. The Sauk II–III sequence boundary falls within the shaded box (see text). All isotope measurements in the Worm Creek interval represent micritic patches within sandy carbonates. Inset showing the progradation of the Worm Creek Quartzite is based on a measured section ~100 km to the north in the Fish Creek Basin of southeastern Idaho (near Grace) with additional control points taken from Haynie (1957), Armstrong (1969), and Wakeley (1975). *Crepicephalus* Zone biostratigraphic control point in northern extent of Worm Creek Quartzite (shown in inset) is based on an unpublished collection made by A.R. Palmer (ICS-1470).

eter. The analytical precision based on duplicate analyses and on multiple analyses of NBS19 was ≤ 0.04‰.

Sequence Stratigraphy, Biostratigraphy, and $\delta^{13}C$ Trends

Smithfield Canyon Section, Northern Utah.—A well-exposed Marjuman–Steptoean succession examined in the Bear River Range of northern Utah represents carbonate platform deposition seaward of the tectonic hinge zone but proximal to a source of terrigenous, inner-detrital-belt sediments (Palmer 1971). The source of siliciclastic material is inferred to be from the northwest in central Idaho, a feature known as the Lemhi Arch, on the basis of thickness and age trends in sandstone units (Palmer 1971). In addition to sampling at Smithfield Canyon for $\delta^{13}C$ stratigraphy, we conducted a detailed decimeter-scale facies analysis and collected trilobites from fourteen horizons (Fig. 4), because no previous study of this locality has been published. New trilobite collections reveal faunas diagnostic of the *Crepicephalus* (late Marjuman), *Aphelaspis, Dunderbergia,* and *Elvinia* zones and allow dating of important lithofacies changes.

The uppermost part of the *Crepicephalus* Zone at Smithfield Canyon is a digitate biostrome complex composed of two stacked microbialite horizons each 3 m thick. The top 60 cm of the biostrome complex, which contains the Marjuman–Steptoean boundary (Fig. 4), is made up of beds of ribbon limestone and microbialite that are cut by numerous, sharp, planar truncation surfaces. Abundant vertical spar-filled tubes (?borings) 2–3 mm in diameter penetrate 1–5 cm into the beds. Early to middle *Aphelaspis* Zone strata that lie directly above the biostrome complex are dominated by silty to sandy ribbon limestones and edgewise flat-pebble conglomerates. The lowest coarse terrigenous clastics in the Smithfield Canyon section occur in this interval as thin to medium interbeds of quartzose calcarenite and calcareous fine- to medium-grained sandstone displaying transitional hummocky to swaly cross-stratification. These sandy strata underlie a 30-m-thick carbonate succession with an age spanning the upper *Aphelaspis* and lower *Dunderbergia* zones and consisting predominantly of ribbon rock (cf. Cowan and James 1993) with intercalations of calcisiltite to fine peloidal calcarenite locally displaying hummocky cross-stratification and wave-ripple cross-lamination. These packages are interpreted as shelf deposits formed below fair-weather wave base during storms. Also commonly interbedded with ribbon rock are lenses and beds of ooid calcarenite with undulatory, wedge-shaped sets of hummocky to planar-tabular cross-stratification that represent relatively brief episodes of impingement of the fair-weather wave base on the shelf. Lithofacies appear to be arranged in coarsening-upward packages 0.1–5 m thick.

Stratigraphically above this ~30-m-thick, clean subtidal carbonate succession (Nounan Formation; Fig. 4), siliciclastic sand is present over a 5-m interval approaching the contact with the Worm Creek Quartzite Member of the St. Charles Formation (Fig. 4). The Worm Creek Quartzite lies a few meters above the lowest collected *Dunderbergia* trilobites and is by far the thickest and most prominent siliciclastic unit in the Middle to Upper Cambrian succession in the region (Williams and Maxey 1941; Williams 1948; Rigo 1968; Trimble and Carr 1976; Oriel and Platt 1980). The lower contact of the Worm Creek is gradational from relatively sand-free ribbon rock to sandy grainstone to carbonate-cemented sandstone to relatively pure quartz sandstone (Fig. 4). A corresponding change in sedimentary structures shows a transition from small-scale hummocky cross-strata to amalgamated hummocky cross-strata to trough cross-strata. The lower one-half of the Worm Creek Quartzite coarsens upward, culminating in a medium- to coarse-grained sandstone with trough cross-strata and wedge to tabular sets of tangential and planar cross-stratification representing an upper-shoreface setting. *Skolithos* burrows and sharply defined truncation surfaces with centimeter-scale relief and medium- to coarse-grained sandstone intraclasts are common throughout this interval, which we interpret to reflect the Sauk II–III sequence boundary zone. The upper half of the Worm Creek mirrors the lower half, with medium- to coarse-grained cross-stratified sandstone passing transitionally upward to finer-grained and progressively more carbonate-rich beds with hummocky cross-strata. An *Elvinia* Zone fauna occurs a few meters above the top of the Worm Creek in a succession of ribbon rock cycles (Fig. 4) and brackets the quartzite biostratigraphically to the mid-Steptoean Sauk II–III interval as recognized elsewhere in Laurentia.

Carbon isotope trends show the SPICE excursion through 80 m of section included within the upper Nounan and lower St. Charles formations (Fig. 4). The beginning of the SPICE is difficult to define precisely in this section (compare with Montañez et al. 2000), but a shift from baseline values to a consistently positive trend appears to occur a few meters above the base of the Steptoean, corresponding to the lowest stratigraphic occurrence of beds of quartzose sandstone (Fig. 4). Peak $\delta^{13}C$ values near + 4.7‰ are observed over a broad interval, which includes both the carbonates of the upper Nounan and the Worm Creek. The $\delta^{13}C$ peak is similar in magnitude to that observed elsewhere in Laurentia, indicating nearly complete preservation of the SPICE (Saltzman et al. 2000). Values begin to decrease towards a baseline near 0‰ just above the sandstone intraclast bed in the middle of the Worm Creek.

Central Iowa Subsurface Section.—The Upper Cambrian succession examined in the Rhinehart A-1 corehole in central Iowa represents deposition on a slowly subsiding shelf (the Hollandale ''Embayment'') in the cratonic interior (Runkel et al. 1998). The Steptoean is ~20% as thick here as the Smithfield Canyon section and nearly 10% as thick as sections in eastern Nevada (Saltzman et al. 1998). The Rhinehart core was studied previously as part of a detailed biostratigraphic and sedimentologic analysis of the Upper Mississippi Valley region by Runkel et al. (1998). Late Marjuman facies in central Iowa consist mainly of carbonate mudstones and wackestones with minor shale intervals and abundant glauconite (Fig. 5), reflecting slow accumulation in an offshore setting below fair-weather wave base. An upward-coarsening sequence that begins a few centimeters above the basal *Aphelaspis* Zone extinction horizon in the Rhinehart core coincides with a regional-scale basinward shift in siliciclastic-dominated facies belts recorded farther inboard on the craton (Fig. 5; Runkel et al. 1998). This overall Steptoean shallowing culminates in 2.5 m of sandy, cross-stratified trilobite grainstones near the top of the *Dunderbergia* Zone. The Sauk II–III boundary was placed by Runkel et al. (1998) within this grainstone unit because the medium- to coarse-grained, rounded quartz sand was presumed to have been derived from the encroaching cratonal shoreline and because it separates *Aphelaspis*- and *Dunderbergia*-age regressively stacked facies below from *Elvinia*-age transgressively stacked facies above. Above the Sauk II–III boundary, mud-rich facies are dominant in the *Elvinia* Zone in the Rhinehart core, recording water depths that appear similar to those in the underlying Marjuman Stage.

The $\delta^{13}C$ stratigraphy of the Rhinehart core (Fig. 5) delineates a remarkably well-preserved SPICE excursion despite the highly condensed nature of the Steptoean sequence. As in the Smithfield Canyon section, $\delta^{13}C$ rises from 0 to + 1‰ in the interval immediately preceding the mass-extinction horizon at the base of Steptoean Stage, beginning a sharp increase in $\delta^{13}C$ in the lower part of the *Aphelaspis* Zone that reaches a peak (+ 3.8‰) within the cross-stratified grainstones of the *Dunderbergia* Zone. A nearly symmetrical return to pre-excursion values close to 0‰ occurs in the lower part of the *Elvinia* Zone (Fig. 5).

Felix Cove Section, Western Newfoundland.—The Marjuman–Steptoean succession in western Newfoundland is well exposed in gently dipping strata in sea cliffs along the south coast of the Port au Port Peninsula. The Cambrian–Ordovician succession here has been interpreted to represent deposition on a paleosouth-facing continental shelf bordering the Iapetus Ocean (James et al. 1989). The studied Marjuman–Steptoean interval, which straddles the Felix Cove and Man O'War members of the Petit Jardin Formation (Fig. 6), is relatively thin compared to sections in the Great Basin and likely represents deposition landward of the tectonic hinge zone or within the inner flexural wedge (Chow and James 1987b; James et al. 1989). As was the case with the Rhinehart core in central Iowa, a detailed, regional sedimentologic and biostratigraphic framework is available for the Felix Cove section (Chow and James 1987a, 1987b; Cowan and James 1992, 1993; Westrop 1992).

Upper Marjuman and lower Steptoean carbonates of the Felix Member (Fig. 6) are composed mainly of ooid–peloid grainstones with abundant dolomicrite as flasers, lenses, interbeds, and intraclasts (i.e., the ''unwashed oolite'' of Cowan and James 1993). Less abundant lithofacies include microbialite mounds as decimeter- to meter-scale bioherms to laterally linked biostromes (stromatolitic and/or thrombolitic textures), ooid grainstone, and ribbon limestone. Medium- to coarse-grained, frosted, well-rounded quartz sand first appears stratigraphically as scattered grains and millimeter-thick stringers in oolite strata of uppermost *Crepicephalus* Zone age (Fig. 6; just above *Terranovella dorsalis* and *Crepicephalus iowensis* in the FC1 and FC2 collections of Westrop 1992). More prominently, however, a bed of quartz sandstone ~1 m thick overlies a microbialite biostrome that yielded a trilobite collection containing *Dytremacephalus strictus* and suggests a maximum age for the apex of terrigenous influx of middle to late *Aphelaspis* Zone (Westrop 1992). The sandstone unit (Fig. 6) is likely an expression of the Sauk II–III regression (Chow and James 1987b; James et al. 1989) and has been attributed to exposure of inboard platform areas and eolian influx of sand across desiccated tidal flats (Cowan and James 1993). Approximately 8 m above the quartz sandstone, a trilobite collection in unwashed ooid–peloid grainstone yields an *Elvinia* Zone fauna (A.R. Palmer, Institute for Cambrian Studies collection ICS-1406), indicating that strata enclosing the Sauk II–III boundary are anomalously thin. The overlying, deeper-water lithofacies of the Man O'War Member yield *Elvinia* Zone trilobites ~10 m above the base and consist of shaly to clean ribbon limestones and dolostones, with rare beds of ooid to pisoid grainstone and large (up to ~5 m tall) microbialite biostromes and bioherms (Kennard and James 1986). The contact between the Felix Member and the overlying Man O'War member would thus appear to be an expression of the *Elvinia* Zone drowning commonly observed throughout Laurentia (Osleger 1995).

Carbon isotope results from the Felix Cove section record the SPICE excursion in ~20 m of the upper Felix and lower Man O'War members (Fig. 6). Values rise from near −1‰ to 0‰ in the upper part of the

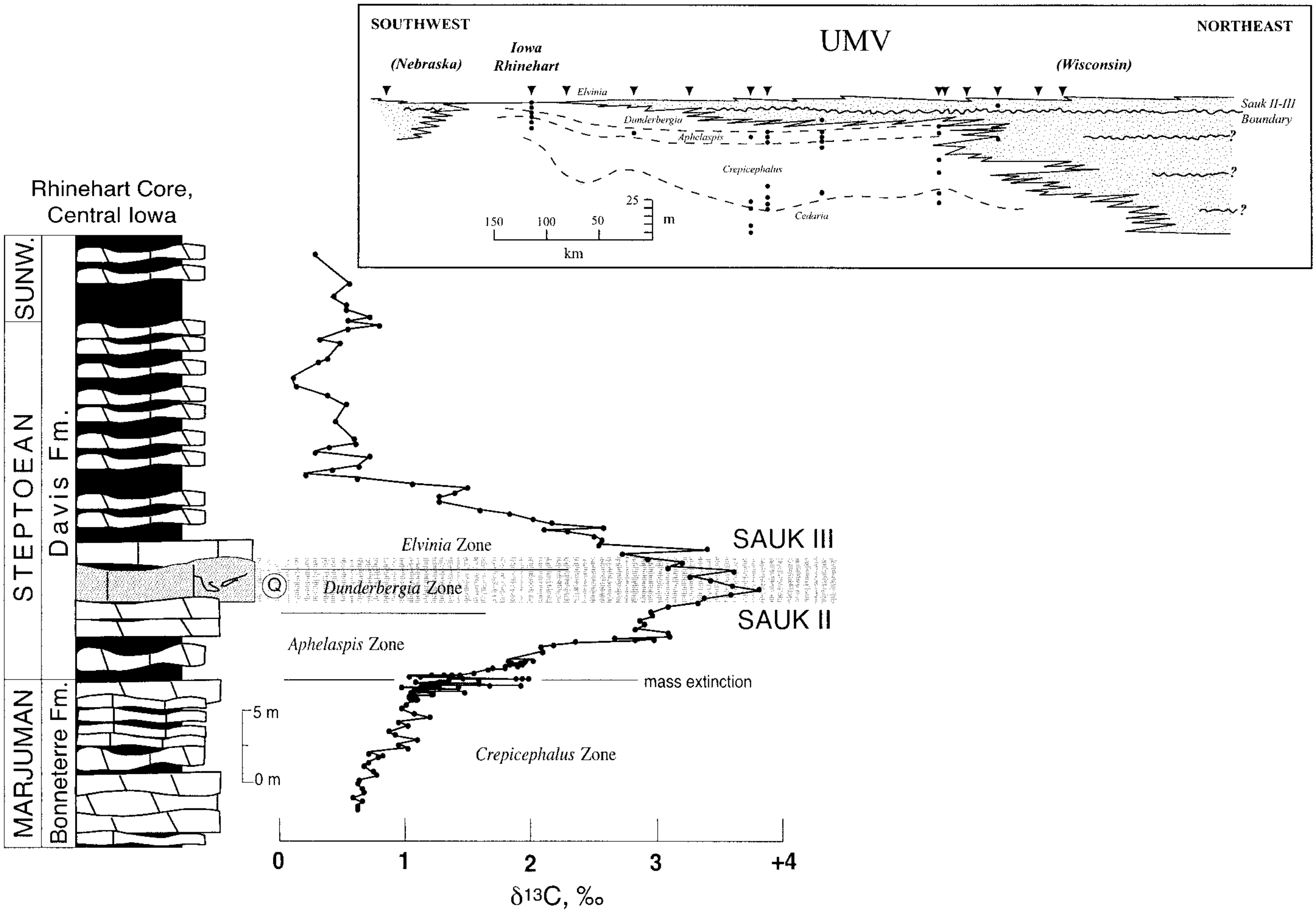

FIG. 5.—Carbon isotope stratigraphy of the Rhinehart A-1 core drilled in Dallas County, central Iowa. Trilobite zones and measured section are after Runkel et al. (1998). See Legend in Figure 7. Shaded box in the upper part of the *Dunderbergia* Zone represents sandy interval that marks maximum regression associated with the Sauk II–Sauk III subsequence boundary. Inset showing progradational Wonewoc Sandstone straddling the Sauk II–Sauk III boundary is based on data in Runkel et al. (1998). The Wonewoc is sandwiched by fine siliciclastic and carbonate facies deposited in offshore shelf environments.

Marjuman and reach peak $\delta^{13}C$ values of + 2.2‰ just above the quartz sandstone bed interpreted to mark the Sauk II–Sauk III sequence boundary in this and previous studies. Although the SPICE is clearly present at Felix Cove, the $\delta^{13}C$ values in the Marjuman and Steptoean are about 2‰ lower than values recorded in Iowa, Utah, and other continents, perhaps reflecting local modification of the global seawater $\delta^{13}C$ signal (cf. Patterson and Walter 1994; Holmden et al. 1998).

DISCUSSION

The carbon isotope results from three Laurentian basins provide a test of the relative synchroneity of regressive sequences that have been attributed to the craton-wide Sauk II–III regression (Fig. 7), independent of traditionally used biozones, and also have paleoceanographic implications related to the partitioning of oxidized and reduced forms of carbon. The discussion is therefore divided into two sections. The first part addresses the correlation of the SPICE with a major regressive–transgressive event recorded in strata throughout Laurentia. Secondly, we consider causal connections between the two events in relation to paleogeographic constraints on carbon burial and weathering during the Steptoean.

Sea-Level Changes during the SPICE $\delta^{13}C$ Excursion in Laurentia

The Upper Cambrian of the Upper Mississippi Valley is a key region for documenting eustatic signals because it contains a record of shoreline migration in a tectonically stable cratonic shelf. In the Rhinehart core, the rising limb of the SPICE tracks a shallowing-upward carbonate sequence that culminates in ~2.5 m of sandy, cross-stratified grainstones chosen by Runkel et al. (1998) to mark the Sauk II–III boundary (Fig. 5). The late Sauk II regressive event is expressed in the siliciclastic succession as progradational, offlapping parasequence sets (cf. Runkel 1994; Runkel et al. 1998; Runkel et al. 1999; Tape et al. 2003). The succession is capped by a subtle sequence-bounding unconformity that has been tracked into the subsurface from the Inner Detrital Belt to settings transitional with the Middle Carbonate Belt, including our key section in central Iowa, using numerous cores and borehole geophysical logs (Fig. 5; Runkel et al. 1998). The maximum seaward extent of the shoreline occurred in late *Dunderbergia* to earliest *Elvinia* zone time, marked by the appearance of coarse siliciclastics in otherwise carbonate-dominated sections from central Iowa to Missouri. Subsequent transgression (*Elvinia* zone time) is recorded by landward stacking of facies that culminate in the retreat of the shoreline far landward of the presently known extent of Cambrian strata in the Upper Mississippi Valley (Runkel et al. 1998).

The Smithfield Canyon section, interpreted within the context of the regional stratigraphy of equivalent strata in Utah and Idaho (Fig. 4), records a history similar to that of the Upper Mississippi Valley (Fig. 5). The increase in age and landward thickening of the lower part of the Worm Creek Quartzite to the north into Idaho reflects the seaward progradation

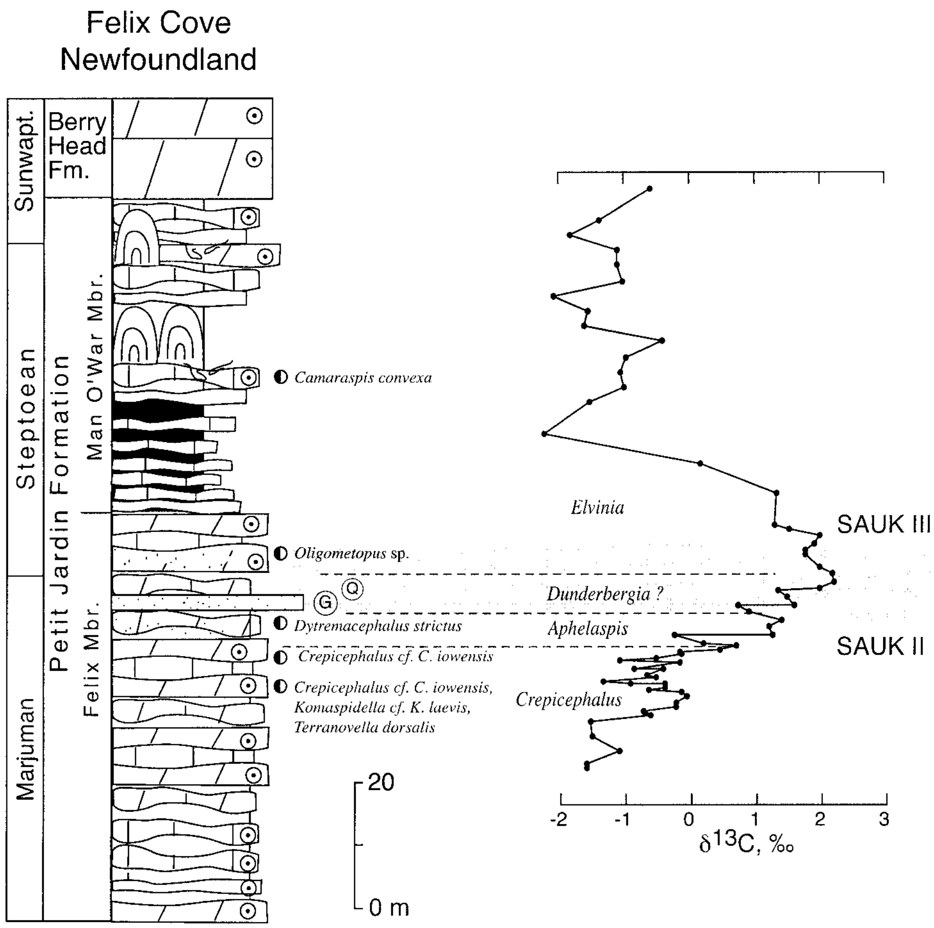

FIG. 6.—Carbon isotope stratigraphy from Felix Cove in western Newfoundland. Trilobite collections are from Westrop (1992) and A.R. Palmer (unpublished). Measured section is after Cowan and James (1993). See legend in Figure 7. Shaded box represents Sauk II–III sequence boundary zone (see text).

of a siliciclastic shoreline in the *Aphelaspis* and *Dunderbergia* zones. The siliciclastic source in this area appears to have been relatively mud-poor, and incursions of quartzose sands were not accompanied by commensurate increases in the shaliness of the fine fraction in the ribbon limestones in this succession. At Smithfield Canyon, the lowest occurrence of terrigenous clastics may represent an early record of this shoreline regression onto the carbonate platform, which occurred in the early to middle *Aphelaspis* Zone coincident with what we interpret to be the onset of the SPICE. Although paleobathymetric changes in the carbonate facies that record the rising limb of the SPICE are subtle, the incremental increase in oolite approaching the Worm Creek Quartzite is interpreted to indicate a shallowing that kept the subtidal carbonate facies belt in more agitated conditions. Carbon isotope values reach $\sim+4.5$‰ within this regressive carbonate sequence and remain high well into the Worm Creek. Maximum regression at Smithfield Canyon is marked by an influx of quartz sand (rather than by a karst surface on top of the carbonate succession), and the medium- to coarse-grained intraclastic sandstone interval that marks the Sauk II–III sequence boundary (cf. Runkel et al. 1998) within the Worm Creek coincides with the decrease from peak δ^{13}C values of the SPICE. Thus, from a regional perspective, the rising limb of the SPICE in the Utah–Idaho area corresponds to regional progradation of the Worm Creek shoreface, with the reversal in the isotope values coinciding with transgression (Fig. 4). Support for a link between relative sea level and δ^{13}C during late Marjuman and early Steptoean time can be seen in the close correspondence between carbon isotope shifts and shoreline movements in the cratonic interior (Fig. 8).

The record at Felix Cove in western Newfoundland is similar to Smithfield Canyon in that carbonate facies associations and inferred changes in water depth were subtle during the Steptoean, and it is the influx of quartz sand in the context of the regional sequence stratigraphic framework (Chow and James 1986b; Cowan and James 1993; cf. Osleger and Montañez 1996) that suggests a sea-level drop during the rising limb of the SPICE (Fig. 6). The Upper Cambrian succession on the Port au Port Peninsula is essentially devoid of quartz sand (Fig. 6), with the exception of the prominent sandy interval in the Felix Member (Chow and James 1987b; Cowan and James 1993). The sub-*Elvinia* Steptoean succession in this region has long been recognized as anomalously thin (James et al. 1989). Because the SPICE peak of $\sim+2$‰ is several per mil below that found elsewhere, the observed stratigraphic thinning is consistent with erosion or nondeposition. However, it is also possible that the local oceanographic setting in western Newfoundland, where δ^{13}C values for the Upper Cambrian as a whole are low compared with other regions (Saltzman et al. 2000), may have reduced (rather than truncated) the peak of the SPICE at Felix Cove section as a result of semi-restriction and increased input of isotopically light carbon into the water mass.

Regression at the Sauk II–III Boundary: A Eustatic Drop?

Our combined results from three widely spaced sections in northern Laurentia provide strong support for the interpretation that Sauk II–III strata record a major, continental-scale regressive–transgressive event. Although eustatic changes in sea level cannot be established as the driving mechanism for this event with complete confidence, our data, as well as those in equivalent sections on other continents, are compatible with such an interpretation. Support for a eustatic drop at the Sauk II–III boundary may be found in correlative intervals of seamount-margin collapse in Kazakhstan (Cook et al. 1991), carbonate-platform exposure in Queensland, Australia

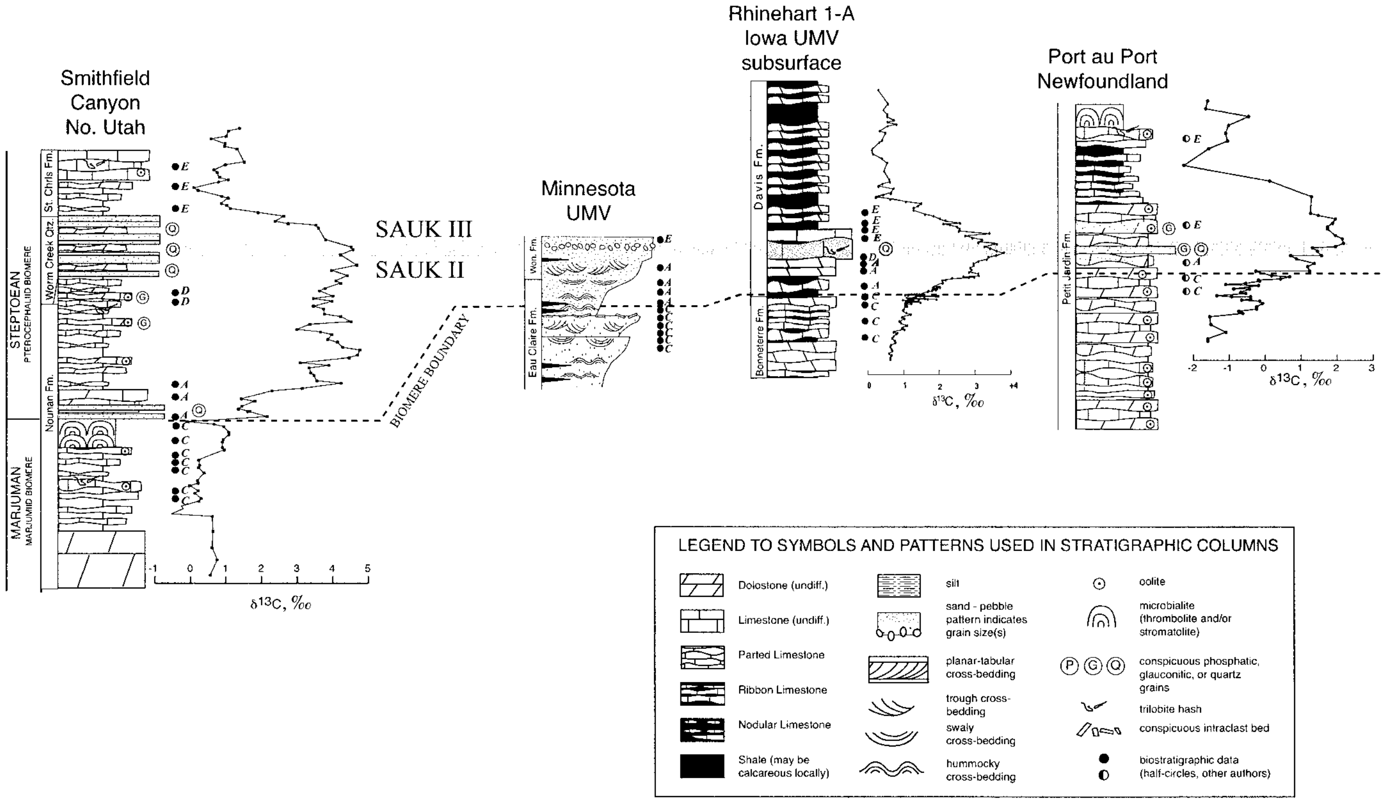

FIG. 7.—Stratigraphic cross section showing carbonate sections analyzed and their correlation with siliciclastic paleoshoreline facies of Minnesota (Runkel et al. 1998; Runkel et al. 2000). Position of the Sauk II–III boundary chosen on lithologic evidence. C = *Crepicephalus* Zone, A = *Aphelapsis* Zone, D = *Dunderbergia* Zone, E = *Elvinia* Zone.

(Henderson 1976), and an anomalously thin Steptoean-equivalent section on the Shantung Peninsula of China (A.R. Palmer, unpublished data).

The duration of the sea-level changes and associated SPICE event is constrained by two U–Pb dates—one (491 ± 1 Ma; Davidek et al. 1998) from the *Peltura scarabaeoides* zone in Welsh Avalonia (= *Eurekia apopsis* zone of Laurentia) and the other (503 ± 4 Ma; Perkins and Walshe 1993) from roughly the *Lejopyge laevigata* level in Tasmania (= early *Cedaria* zone of Laurentia; Shergold 1997). Assuming that there are approximately 15 trilobite zones between these two U–Pb ages in most parts of the world, each zone, on average, lasted about 0.8 My. Given that the SPICE excursion persisted for 4.5 zones (*Aphelaspis* through mid-*Elvinia* of Palmer 1965), its duration was about 3.6 My, with the increase and decrease in $\delta^{13}C$ lasting ~2.8 and ~0.8 My, respectively.

On time scales of several million years or less, glacio-eustasy provides the simplest mechanism for the Sauk II–III regressive–transgressive couplet. However, given the lack of evidence for high-latitude glaciation of Baltica during the Steptoean, we must examine other causes. The apparently small rate of change (a few tens of meters in a few million years) of sea-level fall is consistent with changes in seafloor spreading rates, but such controls typically operate over much longer time scales (Pitman 1978). The long response times of volumetric changes in the ridge system are related to the thermal origin of ridge elevation (with a thermal time constant of about 65 My). For example, Dewey and Pitman (1998) calculated a 2.93 m/My rate of sea-level drop as a result of changes in the volume of the mid-ocean ridge system averaged between 80 and 10 Ma (with a total magnitude of the sea level fall of 205 m), and a 2.06 m/My rate of sea level rise from 205 Ma (Hettangian) to 95 Ma (Cenomanian) (although see Rowley 2002). Thus, although the relatively abrupt elimination of a ridge segment could have interrupted the long-term Sauk transgression enough to cause the Sauk II–III subsequence boundary, the generally slow tectonoeustatic mechanisms that alter the volume of the ocean basins (including orogeny and crustal shortening) are unlikely candidates.

Transfer of ocean water into various land reservoirs (lakes and wetlands) under cooler climatic conditions (Jacobs and Sahagian 1993) or the buildup and melting of alpine glaciers could produce small, short time eustatic fluctuations (Fischer and Hinnov 1997). However, Dewey and Pitman (1998) calculated that increasing the volume of present-day alpine glaciers by a factor of ten would change sea level by only 2.4 m, with a factor-of-fifty change needed to change sea level by 12 m or more. Heating and cooling of the oceans is another mechanism that can have a small but measurable effect on sea level. Annual sea-level cycles with an amplitude of ~0.1 m are produced by temperature fluctuations of 0.2°C in the modern ocean (ATOC Consortium 1998), and thus reducing the whole-ocean temperature by 5°C could potentially lower sea level by several meters. Although such an effect may be observable on a nearly subhorizontal Cambrian cratonal platform, a temperature drop of this magnitude—which is on the scale of the Eocene–Oligocene boundary (~35 to 33 Ma) cooling step (Zachos et al. 2001)—seems unlikely inasmuch as no faunal extinctions occur at the peak of the SPICE (only at the biomere boundaries above and below). Even if several meters of sea-level fall can be attributed to cooling induced by a reduced greenhouse effect, the remaining ≥ 10 m of sea-level fall during the Sauk II–III regression must be accounted for elsewhere.

Large-scale ''regional'' mechanisms for sea-level change, such as in-plane stress, may have been important during the Steptoean, and can produce relatively rapid sea-level changes of 10–50 m (Dewey and Pitman 1998). Sabadini et al. (1990) demonstrated that rapid true-polar-wander episodes of about 1° per million years can also produce sea-level changes on the order of 20–50 meters over 1 million years. Such mechanisms allow predictions about the differential response of sea level in basins that lie on

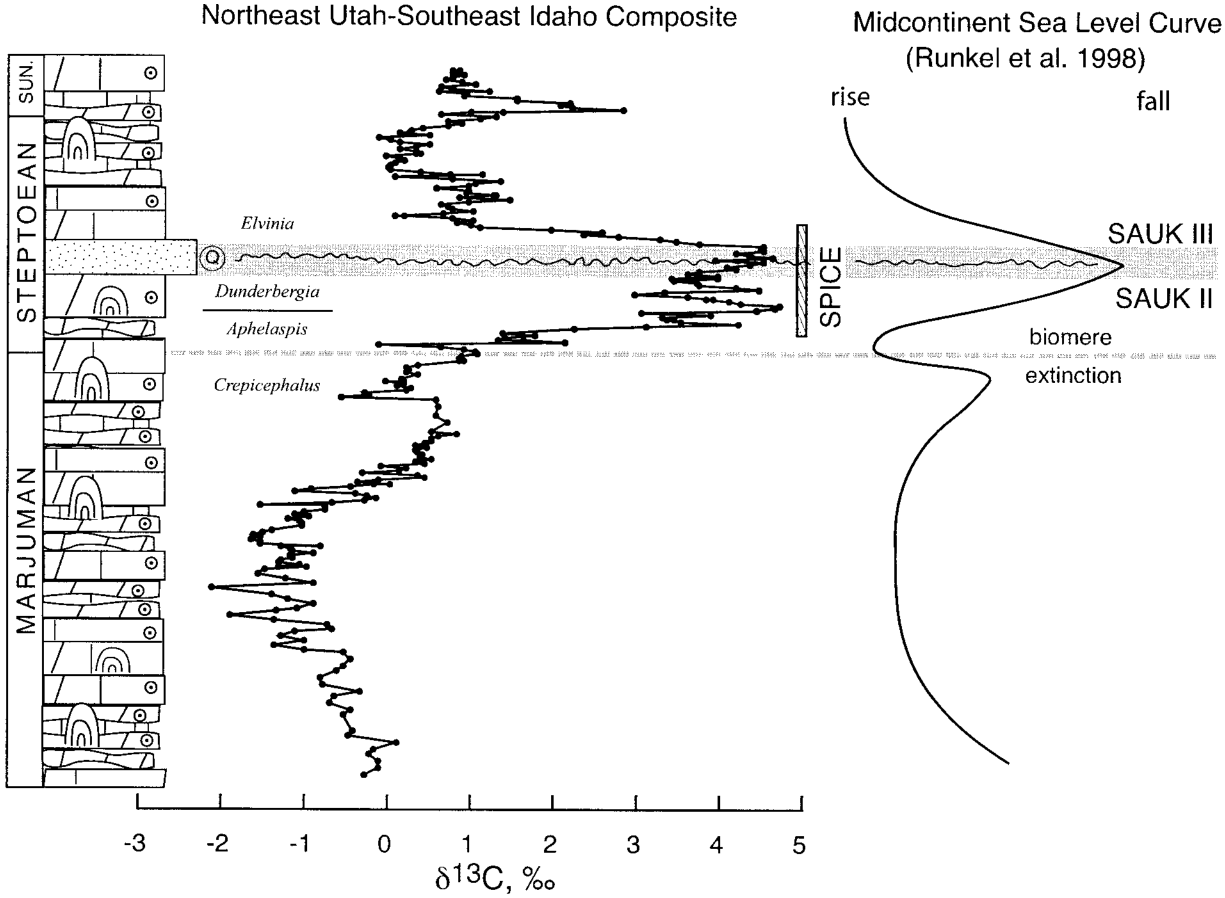

FIG. 8.—Carbonate $\delta^{13}C$ composite from Smithfield Canyon, northern Utah (Fig. 4; late Marjuman through early Sunwaptan) and a lower to middle Marjuman section ~100 km to the north in the Fish Creek Basin of southeastern Idaho (near Grace; inset Fig. 4). Upper Mississippi Valley relative sea-level curve is from Runkel et al. (1998).

opposite sides of the continent or globe that can potentially be tested as the sequence stratigraphic record of the SPICE interval is more thoroughly documented globally.

SPICE Mechanisms: Organic Matter Burial versus Weathering

The close coincidence in time between the peak of SPICE and maximum regression may indicate a causal connection. Excursions in the $\delta^{13}C$ of marine carbonates on time scales longer than $\sim 10^5$ years result from changes in the fraction of carbon buried as organic matter or changes in rocks containing organic carbon or carbonate carbon that are exposed to weathering. In addition, the sensitivity of the photosynthetic carbon isotope effect to changes in pCO_2 can modify the seawater $\delta^{13}C$ signal (Kump and Arthur 1999). The steady-state isotopic mass-balance equation traditionally used to interpret $\delta^{13}C$ excursions is

$$f_{org} = (\delta_w - \delta_{carb})/\Delta_B.$$

The fraction of carbon buried as organic matter (f_{org}) is calculated from the $\delta^{13}C$ of marine carbonates (δ_{carb}) by assuming constant $\delta^{13}C$ values for the riverine carbon flux (δ_w) and the isotopic difference between carbonates and organic matter (Δ_B) (Kump and Arthur 1999). Using the Phanerozoic average values of $\delta_w = -5‰$ and constant $\Delta_B = 30‰$ from Kump and Arthur (1999), a shift from 0‰ to + 5‰ during the SPICE corresponds to a doubling of f_{org}. Simulations using the one-dimensional ocean model of Junge et al. (1975) and Broecker and Peng (1982) as developed by Kump (1991) yield similar results. We can further calculate that over 10^{19} moles of excess carbon was buried (above and beyond steady state) during the SPICE interval using 5×10^{19} moles C/My for the long-term riverine input (Kump and Arthur 1999) and then doubling the fraction of this total C that was buried as organic matter (f_{org}) for 3 My (also assuming constant Δ_B). A + 5‰ $\delta^{13}C$ shift could alternatively be explained by an increase in carbonate weathering by ~70% to 90% that increased the $\delta^{13}C$ of the riverine flux (δ_w). However, such a change is unlikely without significant ice-sheet buildup over silicate terranes (containing organic matter) at high latitudes, and thus excess burial of organic matter is assumed to play the major role in the SPICE.

The burial of organic matter may increase and be sustained on a million-year time scale in response to (1) a greater flux of phosphate from the continents, or possibly upwelling from the deep ocean, or (2) an increase in the carbon-to-phosphorus ratio of the organic matter that is buried (Kump and Arthur 1999; Schrag et al. 2002). While it is possible that phosphate-rich deposits of Late Proterozoic and Early Cambrian age were preferentially exposed and eroded during the Sauk II–III regression, such deposits are typically associated with large amounts of ^{12}C-enriched organic matter that would lower riverine $\delta^{13}C$. If enhanced production combined with preferential regeneration of phosphorus relative to carbon in organic matter underlying oxygen-poor waters (cf. Van Cappellen and Ingall 1994) caused the SPICE, where did such environments exist?

In the modern ocean, the burial of organic matter occurs predominantly in fine-grained siliciclastic sediments in deltaic, upper shelf, and slope settings where high rates of sedimentation, nutrient fluxes, and suspended loads of terrestrial organic matter are found (Berner 1982; Hedges and Keil

1995). Berner (1982) showed that the organic carbon burial rate in these environments today (~70–80% of total burial) is far greater than that calculated for biogenous sediments underlying regions of high productivity, shallow-water carbonate settings, or remaining pelagic sediments of low-productivity regions. Recognition that the adsorption of carbon compounds onto clay-mineral surfaces plays an important role in preservation of organic carbon (Kennedy et al. 2002) further underscores the potential importance of fine-grained siliciclastic deposits in deltaic, upper shelf, and slope settings as short-term sinks for buried organic matter. It is therefore reasonable to look to these depositional environments for the excess organic matter indicated by the SPICE, although the lack of an extensive terrestrial biosphere (and thus negligible suspended loads of organic matter) in the early Paleozoic may have shifted organic-carbon burial to other settings to some degree.

Schrag et al. (2002) recently proposed that the high $\delta^{13}C$ values in the Neoproterozoic were linked to extensive local anoxia in fine-grained siliciclastic environments of high rates of organic-carbon burial in tropical river systems. This resulted in more efficient phosphate recycling and higher carbon-to-phosphorus ratios of the buried organic matter. According to this model, the ultimate driver for the positive $\delta^{13}C$ excursions in the Neoproterozoic was the relatively high percentage of continental area in the tropics at this time, which produced maximal overlap of zones of high productivity (upwelling) and regions of high riverine phosphorus and sediment input. Although the time scale of the SPICE excursion may be less than the Neoproterozoic shifts, continental positioning during the Steptoean may have remained favorable for such a scenario. The trigger may have been the Sauk II–III regression, which exposed fresh basement rock in the interiors of cratons to weathering and produced an expanded area of fine-grained siliciclastic environments. Such deposits extend regionally across the Upper Mississippi Valley during the SPICE interval (latest *Crepicephalus* through *Dunderbergia* zones) and overlie sandstone- and carbonate-dominated successions across much of the subsurface (Howe et al. 1972; McKay 1988; Kurtz 1989; Runkel et al. 1998). Similar siliciclastic shelf environments that may have existed in the Late Cambrian of Australia and Antarctica (e.g., Shergold et al. 1985) and parts of the Middle East and central Asia during the SPICE are more difficult to date because of a lack of fossil control.

Significant amounts of organic matter could also have been buried in the deep ocean, which may have been anoxic in the early Paleozoic (Wilde and Berry 1984). An expanded area of deeper-water ''black shale'' environments, analogous to those associated with Mesozoic oceanic anoxic events (Arthur et al. 1987), could also have been facilitated by organic-matter redeposition from shallow shelf regions during the Sauk II–III regression. The vast areas of shallow-water carbonate production that remained submerged during the Steptoean should also not be overlooked as the sites of significant burial of organic matter during the SPICE if original TOC values were as high as those documented in the fine-grained carbonate muds deposited in intraplatform basins and canyons of the Bahamas (Crevello et al. 1984). However, postdepositional effects that appear to systematically lower TOC values in ancient limestones relative to shales (Gehman 1962) make it difficult to document such a repository. Regardless of where enhanced organic-carbon burial is hypothesized to have taken place, any model for the coincidence of the SPICE and Sauk II–III regression must also factor in the potential for changes in riverine $\delta^{13}C$ caused by the erosion of newly exposed carbonate- or organic-rich deposits elsewhere.

CONCLUSIONS

Carbon isotope data from Upper Cambrian sections in northern Utah, central Iowa, and western Newfoundland record the globally recognized SPICE $\delta^{13}C$ event of up to + 5‰. Peak $\delta^{13}C$ ratios are well correlated by trilobite biostratigraphy to the middle of the Steptoean Stage (*Dunderbergia* Zone) and track the regression associated with the Sauk II–III boundary. Regression is indicated by influxes of quartz sand into carbonate-platform settings in all three widely separated basins and by shallowing of carbonate environments in central Iowa and Utah. In the stratigraphic successions of this and previous investigations, the position of the Sauk II–III boundary based on lithologic indicators can be used to predict the peak interval of the SPICE remarkably well and is consistent with synchroneity of facies shifts across Laurentia controlled by eustasy. The observed pattern of sea-level fall and peak $\delta^{13}C$ values may indicate that the SPICE is similar to events recognized in the Late Ordovician (e.g., Kump et al. 1999) and Silurian (e.g., Wenzel and Joachimski 1996; Azmy et al. 1998; Saltzman 2001, 2002). An increase in weathering rates during exposure and erosion of the cratonal interiors likely increased the percentage of total carbon buried as organic matter in fine-grained siliciclastic deposits in shelf and upper slope environments during the Steptoean. Increased carbonate weathering relative to terrigenous weathering was likely less of a factor because of the lack of known ice sheets that could have covered up silicate terranes containing organic matter at high latitudes.

ACKNOWLEDGMENTS

R.M. McKay and the Iowa Geological Survey provided generous access to the Rhinehart A-1 core and to unpublished information. Stable-isotope analyses on the Rhinehart core were carried out in the Stable Isotope Laboratories of the University of California, Santa Cruz (J.C. Zachos, PI) and the University of Michigan (K.C. Lohmann, PI) with funds made available by the Institute of Geophysics and Planetary Physics, UCLA. The Smithfield and Felix Cove samples were analyzed at Michigan and the University of Iowa (Luis A. González, PI), respectively. The Henrickson and Berstein Funds of the Carleton College Geology Department defrayed the cost of fieldwork (CAC). We thank C.H. House and M.J. Kennedy for helpful discussion of early versions of the manuscript and B. Oaks and P.K. Link for field advice and assistance. We thank Loren Babcock for discussion of Cambrian chronostratigraphy. This manuscript was greatly improved by the comments of JSR reviewers Ethan Grossman and Michael Joachimski, along with AE Tim Lyons and David Budd.

REFERENCES

ALGEO, T.J., WILKINSON, B.H., AND LOHMANN, K.C., 1992, Meteoric-burial diagenesis of Middle Pennsylvanian limestones in the Orogrande Basin, New Mexico: Water/rock interactions and basin geothermics: Journal of Sedimentary Petrology, v. 62, p. 652–670.

ARMSTRONG, F.C., 1969, Geologic map of the Soda Springs Quadrangle, southeastern Idaho: U.S. Geological Survey, Miscellaneous Geological Investigations, Map I-557.

ARTHUR, M.A., SCHLANGER, S.O., AND JENKYNS, H.C., 1987, The Cenomanian–Turonian oceanic anoxic event II. Palaeoceanographic controls on organic-matter production and preservation, *in* Brooks, J., and Fleet, A. J., eds., Marine Petroleum Source Rocks: Geological Society of London, Special Publication 26, p. 401–420.

ATOC CONSORTIUM, 1998, Ocean climate change: Comparison of acoustic tomography, satellite altimetry, and modeling: Science, v. 281, p. 1327–1332.

AZMY, K., VEIZER, J., BASSETT, M.G., AND COPPER, P., 1998, Oxygen and carbon isotopic composition of Silurian brachiopods: Implications for coeval seawater and glaciations: Geological Society of America, Bulletin, v. 110, p. 1499–1512.

BANNER, J.L., AND HANSON, G.N., 1990, Calculation of simultaneous isotopic and trace element variations during water–rock interaction with applications to carbonate diagenesis: Geochimica et Cosmochimica Acta, v. 54, p. 3123–3137.

BERNER, R.A., 1982, Burial of organic carbon and pyrite sulfur in the modern ocean: Its geochemical and environmental significance: American Journal of Science, v. 282, p. 451–473.

BERNER, R.A., 1990, Atmospheric carbon dioxide levels over Phanerozoic time: Science, v. 249, p. 1382–1386.

BOND, G.C., NICKESON, P.A., AND KOMINZ, M.A., 1984, Breakup of a supercontinent between 625 Ma and 555 Ma: new evidence and implications for continental histories: Earth and Planetary Science Letters, v. 70, p. 325–345.

BRENCHLEY, P.J., MARSHALL, J.D., CARDEN, G.A.F., ROBERTSON, D.B.R., LONG, D.F.G., MEIDLA, T., HINTS, L., AND ANDERSON, T.F., 1994, Bathymetric and isotopic evidence for a short-lived Late Ordovician glaciation in a greenhouse period: Geology, v. 22, p. 295–298.

BROECKER, W.S., AND PENG, T-H., 1982, Tracers in the Sea: Palisades, New York, Eldigio, 690 p.

CHOW, N., AND JAMES, N.P., 1987a, Facies-specific, calcitic and bimineralic ooids from Middle and Upper Cambrian platform carbonates, western Newfoundland, Canada: Journal of Sedimentary Petrology, v. 57, p. 907–921.

CHOW, N., AND JAMES, N.P., 1987b, Cambrian Grand Cycles: northern Appalachian perspective: Geological Society of America, Bulletin, v. 98, p. 418–429.

COOK, H.E., TAYLOR, M.E., ZHEMCHUZHNIKOV, S.V., APOLLONOV, M.K., ERGALIEV, G.K., SARGASKAEV, Z.S., AND DUBININA, S.V., 1991, Comparison of two early Paleozoic carbonate submarine fans—western US and southern Kazakhstan, *in* Cooper, J.D., and Stevens, C.H.,

eds., Paleozoic Paleogeography of the Western United States—II: Los Angeles, SEPM, Pacific Section, p. 847–872.
Cowan, C.A., and James, N.P., 1992, Diastasis cracks—mechanically generated synaeresis-like cracks in Upper Cambrian shallow-water oolite and ribbon carbonates: Sedimentology, v. 39, p. 1101–1118.
Cowan, C.A., and James, N.P., 1993, The interactions of sea-level change, terrigenous-sediment influx, and carbonate productivity as controls on Upper Cambrian Grand Cycles of western Newfoundland, Canada: Geological Society of America, Bulletin, v. 105, p. 1576–1590.
Crevello, P.D., Patton, J.W., Oesleby, T.W., Schlager, W., and Droxler, A., 1984, Source rock potential of Bahamian Trough carbonates, *in* Stow, D.A.V., and Piper, D.J.W., eds., Fine-Grained Sediments: Deep-Water Processes and Facies, London, Blackwell, p. 469–480.
Davidek, K., Landing, E., Bowring, S.A., Westrop, S.R., Rushton, A.W.A., Fortey, R.A., and Adrain, J.M., 1998, New uppermost Cambrian U–Pb date from Avalonian Wales and the age of the Cambrian–Ordovician boundary: Geological Magazine, v. 135, p. 305–309.
Derry, L.A., Kaufman, A.J., and Jacobsen, S.B., 1992, Sedimentary cycling and environmental change in the Late Proterozoic: Evidence from stable and radiogenic isotopes: Geochimica et Cosmochimica Acta, v. 56, p. 1317–1329.
Dewey, J.F., and Pitman, W.C., 1998, Sea-level changes: Mechanisms, magnitudes and rates, *in* Pindell, J.L., and Drake, C., eds., Paleogeographic Evolution and Non-Glacial Eustasy, Northern South America: SEPM, Special Publication 58, p. 1–16.
Finney, S.C., Berry, W.B.N., Cooper, J.D., Ripperdan, R.L., Sweet, W.C., Jacobson, S.R., Soufiane, A., Achab, A., and Noble, P.J., 1999, Late Ordovician mass extinction: A new perspective from stratigraphic sections in central Nevada: Geology, v. 27, p. 215–218.
Fischer, A.G., and Hinnov, L.A., 1997, Orbitally forced glaciation in greenhouse and icehouse times: Part 1. A global model (abstract): Geological Society of America, Abstracts with Programs, Annual Meeting (Salt Lake City), Abstract No. 50204, www.geosociety.org.
Föllmi, K.B., Weissert, H., Bisping, M., and Funk, H., 1994, Phosphogenesis, carbon-isotope stratigraphy, and carbonate-platform evolution along the Lower Cretaceous northern Tethyan margin: Geological Society of America, Bulletin, v. 106, p. 729–746.
Gao, G., and Land, L.S., 1991, Geochemistry of Cambrian–Ordovician Arbuckle Limestone, Oklahoma: Implications for diagenetic $\delta^{18}O$ alteration and secular $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ variation: Geochimica et Cosmochimica Acta, v. 55, p. 2911–2920.
Gehman, H.M., 1962, Organic matter in limestones: Geochimica et Cosmochimica Acta, v. 26, p. 885–897.
Glumac, B., and Spivak-Birndorf, M.L., 2002, Stable isotopes of carbon as an invaluable stratigraphic tool: An example from the Cambrian of the northern Appalachians, U.S.A.: Geology, v. 30, p. 563–566.
Glumac, B., and Walker, K.R., 1998, A Late Cambrian positive carbon-isotope excursion in the southern Appalachians: Relation to biostratigraphy, sequence stratigraphy, environments of deposition, and diagenesis: Journal of Sedimentary Research, v. 68, p. 1212–1222.
Gruszczynski, M., Halas, S., Hoffman, A., and Malkowski, K., 1989, A brachiopod calcite record of the oceanic carbon and oxygen isotope shifts at the Permian/Triassic transition: Nature, v. 337, p. 64–68.
Haynie, A.V., Jr., 1957, The Worm Creek Quartzite Member of the St. Charles Formation, Utah–Idaho [Master's thesis]: Utah State University, Logan, Utah, 39 p.
Hedges, J.I., and Keil, R.G., 1995, Sedimentary organic matter preservation: an assessment and speculative synthesis: Marine Chemistry, v. 49, p. 81–115.
Henderson, R.A., 1976, Stratigraphy of the Georgina Limestone and a revised zonation for the early Upper Cambrian Idamean Stage: Geological Society of Australia, Journal, v. 23, p. 423–433.
Holmden, C., Creaser, R.A., Muehlenbachs, K., Leslie, S.A., and Bergstrom, S.M., 1998, Isotopic evidence for geochemical decoupling between ancient epeiric seas and bordering oceans: Implications for secular curves: Geology, v. 26, p. 567–570.
Howe, W.B., Kurtz, V.E., and Anderson, K.H., 1972, Correlation of Cambrian strata of the Ozark and Upper Mississippi Valley Regions: Missouri Geological Survey and Water Resources Report of Investigations, no. 52, 61 p.
Immenhauser, A., Kenter, J.A.M., Ganssen, G., Bahamonde, J.R., Van Vliet, A., and Saher, M.H., 2002, Origin and significance of isotope shifts in Pennsylvanian carbonates (Asturias, northwest Spain): Journal of Sedimentary Research, v. 72, p. 82–94.
Jacobs, D.K., and Sahagian, D.L., 1993, Climate-induced fluctuations in sea level during non-glacial times: Nature, v. 361, p. 710–712.
James, N.P., Stevens, R.K., Barnes, C.R., and Knight, I., 1989, Evolution of a lower Paleozoic continental-margin carbonate platform, northern Canadian Appalachians, *in* Crevello, P., Wilson, J.L., Sarg, J.F., and Read, J.F., eds., Controls on Carbonate Platform and Basin Development: SEPM, Special Publication 44, p. 123–146.
Joachimski, M.M., and Buggisch, W., 1993, Anoxic events in the late Frasnian—Causes of the Frasnian–Famennian faunal crisis?: Geology, v. 21, p. 675–678.
Joachimski, M.M., Pancost, R.D., Freeman, K.H., Ostertag-Henning, C., and Buggisch, W., 2002, Carbon isotope geochemistry of the Frasnian–Famennian transition: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 181, p. 91–109.
Junge, C.E., Schidlowski, M., Eichmann, R., and Pietrek, H., 1975, Model calculations for the terrestrial carbon cycle: Carbon isotope geochemistry and evolution of photosynthetic oxygen: Journal of Geophysical Research, v. 80, p. 4542–4552.
Kennard, J.M., and James, N.P., 1986, Thrombolites and stromatolites: two distinct types of microbial structures: Palaios, v. 1, p. 492–503.
Kennedy, M.J., Peaver, D.R., and Hill, R.J., 2002, Mineral surface control of organic carbon in black shale: Science, v. 295, p. 657–660.
Kump, L.R., 1991, Interpreting carbo-isotope excursions: Strangelove oceans: Geology, v. 19, p. 299–302.
Kump, L.R., and Arthur, M.A., 1999, Interpreting carbon-isotope excursions: carbonates and organic matter: Chemical Geology, v. 161, p. 181–198.
Kump, L.R., Arthur, M.A., Patzkowsky, M.E., Gibbs, M.T., Pinkus, D.S., and Sheehan, P.M., 1999, A weathering hypothesis for glaciation at high atmospheric pCO_2 during the Late Ordovician: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 152, p. 173–187.
Kurtz, V.E., 1989, Correlation of Upper Cambrian strata in the subsurface of eastern Missouri, *in* Gregg, J.M., Palmer, J.R., and Kurtz, V.E., 1989, Field Guide to the Upper Cambrian of Southeastern Missouri (Stratigraphy, Sedimentology and Economic Geology): Geological Society of America, Field Trip #5 1989 Annual Meeting: Rolla, Missouri, p. 73–83.
Lochman-Balk, C., 1970, Upper Cambrian faunal patterns on the craton: Geological Society of America, Bulletin, v. 81, p. 3197–3224.
Lohmann, K.C., 1988, Geochemical patterns of meteoric diagenetic systems and their application to studies of paleokarst, *in* James, N.P., and Choquette, P.W., eds., Paleokarst: Berlin, Springer-Verlag, p. 58–80.
McKay, R.M., 1988, Stratigraphy and lithofacies of the Dresbachian (Upper Cambrian) Eau Claire Formation in the subsurface of eastern Iowa, *in* Ludvigson, G.A., and Bunker, B.A., eds., New Perspectives on the Paleozoic History of the Upper Mississippi Valley: SEPM, Guidebook, 18th Field Conference of the Great Lakes section, p. 33–53.
Mii, H., Grossman, E.L., and Yancey, T.E., 1999, Carboniferous isotope stratigraphies of North America: Implications for Carboniferous paleoceanography and Mississippian glaciation: Geological Society of America Bulletin, v. 111, p. 960–973.
Montañez, I.P., Banner, J.L., Mack, L.E., Musgrove, M., and Osleger, D.A., 2000, Evolution of the Sr and C isotope composition of Cambrian oceans: GSA Today, v. 10, no. 5, p. 1–7.
Oriel, S.S., and Platt, L.B., 1980, Geological map of the Preston $1^\circ \times 2^\circ$ Quadrangle, southeastern Idaho and western Wyoming: U.S. Geological Survey, Miscellaneous Investigation Series, Map I-1127.
Osleger, D.A., 1995, Depositional sequences on Upper Cambrian carbonate platforms: Variable sedimentologic responses to allogenic forcing, *in* Haq, B.U., ed., Sequence Stratigraphy and Depositional Response to Eustatic, Tectonic and Climatic Forcing: Dordrecht, The Netherlands, Kluwer Academic Publishers, p. 247–276.
Osleger, D.A., and Montañez, I.P., 1996, Cross-platform architecture of a sequence boundary in mixed siliciclastic–carbonate lithofacies, Middle Cambrian, southern Great Basin, USA: Sedimentology, v. 43, p. 197–217.
Osleger, D.A., and Read, J.F., 1993, Comparative analysis of methods used to define eustatic variations in outcrop: Late Cambrian interbasinal sequence development: American Journal of Science, v. 293, p. 157–216.
Palmer, A.R., 1965, Trilobites of the Late Cambrian Pterocephaliid Biomere in the Great Basin, United States: U.S. Geological Survey, Professional Paper 493, 105 p.
Palmer, A.R., 1971, The Cambrian of the Great Basin and adjacent areas, western United States, *in* Holland, H.C., ed., Cambrian of the New World: London, Wiley-Interscience, p. 1–79.
Palmer, A.R., 1981, Subdivision of the Sauk sequence, *in* Taylor, M.E., ed., Short Papers for the Second International Symposium on the Cambrian System: U.S. Geological Survey, Open-File Report 81-743, p. 160–162.
Palmer, A.R., 1998, A proposed nomenclature for stages and series for the Cambrian of Laurentia, *in* Landing, E., ed., Cambrian Subdivisions and Correlations: Canadian Journal of Earth Sciences, v. 35, p. 323–328.
Patterson, W.P., and Walter, L.M., 1994, Depletion of ^{13}C in seawater ΣCO_2 on modern carbonate platforms: Significance for the carbon isotopic record of carbonates: Geology, v. 22, p. 885–888.
Peng, S., Babcock, L.E., and Zhu, M., 2001, Cambrian System of South China, Palaeoworld no. 13, Hefei, University of Science and Technology of China, 310 p.
Perkins, C., and Walshe, J.L., 1993, Geochronology of the Mount Read Volcanics, Tasmania, Australia: Economic Geology, v. 88, p. 1176–1197.
Pitman, W.C., 1978, Relationship between eustasy and stratigraphic sequences of passive margins: Geological Society of America, Bulletin, v. 89, p. 1389–1403.
Railsback, L.B., Holland, S.M., Hunter, D.M., Jordan, M.E., Diaz, J.R., and Crowe, D.E., 2003, Controls on geochemical expression of subaerial exposure in Ordovician limestones from the Nashville Dome, Tennessee, U.S.A.: Journal of Sedimentary Research, v. 73, p. 790–805.
Rigo, R.J., 1968, Middle and Upper Cambrian stratigraphy in the autochthon and allochthon of northern Utah: Brigham Young University, Geology Studies, v. 15, p. 31–66.
Ripperdan, R.L., Magaritz, M., Nicoll, R.S., and Shergold, J.H., 1992, Simultaneous changes in carbon isotopes, sea-level and conodont biozones within the Cambrian–Ordovician boundary interval at Black Mountain, Australia: Geology, v. 20, p. 1039–1042.
Rowley, D.B., 2002, Rate of plate creation and destruction: 180 Ma to present: Geological Society of America, Bulletin, v. 114, p. 927–933.
Runkel, A.C., 1994, Deposition of the uppermost Cambrian (Croixan) Jordan Sandstone, and the nature of the Cambrian–Ordovician boundary in the Upper Mississippi Valley: Geological Society of America, Bulletin, v. 106, p. 492–506.
Runkel, A.C., McKay, R.M., and Palmer, A.R., 1998, Origin of a classic cratonic sheet sandstone: Stratigraphy across the Sauk II–Sauk III boundary in the Upper Mississippi Valley: Geological Society of America, Bulletin, v. 110, p. 188–210.
Runkel, A.C., McKay, R.M., Cowan, C.A., and Saltzman, M.R., 2000, Relative sea level changes across the Marjumiid–Pterocephaliid biomere boundary as recorded in the siliciclastic shoreline strata of the central midcontinent: Stop D *in* 30th Annual Field Conference, SEPM, Great Lakes Section, p. 35–39.
Runkel, A.C., Miller, J.F., McKay, R.M., Shaw, T.H., and Basset, D.J., 1999, Cambrian–Ordovician boundary strata in the central midcontinent of North America: Acta Universitatis Carolinae–Geologica, v. 43, p. 17–20.
Sabadini, R., Doglioni, C., and Yuen, D.A., 1990, Eustatic sea-level fluctuations induced by polar wander: Nature, v. 345, p. 708–710.
Saltzman, M.R., 2001, Silurian $\delta^{13}C$ stratigraphy: A view from North America: Geology, v. 29, p. 671–674.

SALTZMAN, M.R., 2002, Carbon and oxygen isotope stratigraphy of the Lower Mississippian (Kinderhookian–early Osagean), western United States: Implications for seawater chemistry and glaciation: Geological Society of America, Bulletin, v. 114, p. 96–108.

SALTZMAN, M.R., BRASIER, M.D., RIPPERDAN, R.L., ERGALIEV, G.K., LOHMANN, K.C., ROBISON, R.A., CHANG, W.T., PENG, S., AND RUNNEGAR, B., 2000, A global carbon isotope excursion during the Late Cambrian: Relation to trilobite extinctions, organic-matter burial and sea level: Palaeogeography, Palaeoceanography, Palaeoclimatology, v. 162, p. 211–223.

SALTZMAN, M.R., RUNNEGAR, B., AND LOHMANN, K.C., 1998, Carbon-isotope stratigraphy of the Pterocephaliid Biomere in the eastern Great Basin: Record of a global oceanographic event during the Late Cambrian: Geological Society of America, Bulletin, v. 110, p. 285–297.

SCHRAG, D.P., BERNER, R.A., HOFFMAN, P.F., AND HALVERSON, G.P., 2002, On the initiation of a snowball Earth: Geochemistry, Geophysics, Geosystems, v. 3: U1–U21 JUN 27.

SHERGOLD, J.H., 1997, Explanatory notes for the Cambrian correlation chart, *in* Whittington, H.B., and 16 others, Treatise on Invertebrate Paleontology, Part O, Arthropoda 1, Trilobita, revised, vol. 1, Introduction, Order Agnostida, Order Redlichiida: Geological Society of America and University of Kansas, p. 301–311.

SHERGOLD, J.H., JAGO, J.B., COOPER, R.A., AND LAURIE, J.H., 1985, The Cambrian System in Australia, Antarctica and New Zealand, *in* Shergold, J.H., and Palmer, A.R., eds., International Union of Geological Sciences, Publication 19, 85 p.

SLOSS, L.L., 1963, Sequences in the cratonic interior of North America: Geological Society of America, Bulletin, v. 74, p. 93–114.

TAPE, C.H., COWAN, C.A., AND RUNKEL, A.C., 2003, Tidal-bundle sequences in the Jordan Sandstone (Upper Cambrian), southeastern Minnesota, U.S.A.: Evidence for tides along inboard shorelines of the Sauk Epicontinental Sea: Journal of Sedimentary Research, v. 73, p. 354–366.

TRIMBLE, D.E., AND CARR, W.J., 1976, Geology of the Rockland and Arbon Quadrangles, Power County, Idaho: U.S. Geological Survey, Bulletin 1399, p. 1–113.

VAN CAPPELLEN, P., AND INGALL, E.D., 1994, Benthic phosphorus regeneration, net primary production, and ocean anoxia: A model of the coupled marine biogeochemical cycles of carbon and phosphorus: Paleoceanography, v. 9, p. 677–692.

WAKELEY, L.D., 1975, Petrology of the Upper Nounan–Worm Creek Sequence, Upper Cambrian Nounan and St. Charles formations, Southeast Idaho [Master's thesis]: Utah State University, Logan, Utah, 137 p.

WENZEL, B., AND JOACHIMSKI, M.M., 1996, Carbon and oxygen isotopic composition of Silurian brachiopods (Gotland/Sweden); palaeoceanographic implications: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 122, p. 143–166.

WESTROP, S.R., 1992, Upper Cambrian (Marjuman–Steptoean) trilobites from the Port-au-Port Group, western Newfoundland: Journal of Paleontology, v. 66, p. 228–255.

WILDE, P., AND BERRY, W.B.N., 1984, Destabilization of the oceanic density structure and its significance to marine 'extinction' events: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 48, p. 143–162.

WILLIAMS, J.S., 1948, Geology of the Paleozoic rocks, Logan quadrangle, Utah: Geological Society of America, Bulletin, v. 59, p. 1121–1164.

WILLIAMS, J.S., AND MAXEY, G.B., 1941, Cambrian section in the Logan quadrangle, Utah and vicinity: American Journal of Science, v. 239, p. 276–284.

ZACHOS, J., PAGANI, M., SLOAN, L., THOMAS, E., AND BILLUPS, K., 2001, Trends, rhythms, and aberrations in global climate 65 Ma to present: Science, v. 292, p. 686–693.

Received 10 June 2003; accepted 2 December 2003.