Subsidence history of the Ediacaran Johnnie Formation and related strata of southwest Laurentia: Implications for the age and duration of the Shuram isotopic excursion and animal evolution

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ABSTRACT

The Johnnie Formation and associated Ediacaran strata in southwest Laurentia are ~3000 m thick, with a Marinoan cap carbonate sequence at the bottom, and a transition from Ediacaran to Cambrian fauna at the top. About halfway through the sequence, the Shuram negative carbon isotopic excursion occurs within the Rainstorm Member near the top of the Johnnie Formation, followed by a remarkable valley incision event. At its type locality in the northwest Spring Mountains, Nevada, the Johnnie lithostratigraphy consists of three distinctive sand-rich intervals alternating with four siltstone/carbonate-rich intervals, which appear correlative with other regional Johnnie Formation outcrops. Carbon isotope ratios in the sub–Rainstorm Member part of the Johnnie Formation are uniformly positive for at least 400 m below the Shuram excursion and compare well with sub–Shuram excursion profiles from the Khufai Formation in Oman. There is historical consensus that the Johnnie and overlying formations were deposited on a thermally subsiding passive margin. Following previous authors, we used Paleozoic horizons of known biostratigraphic age to define a time-dependent exponential subsidence model, and extrapolated the model back in time to estimate the ages of the Shuram excursion and other prominent Ediacaran horizons. The model suggests that the Shuram excursion occurred from 585 to 579 Ma, and that incision of the Rainstorm Member shelf occurred during the 579 Ma Gaskiers glaciation. It further suggests that the base of the Johnnie Formation is ca. 630 Ma, consistent with the underlying Noonday Formation representing a Marinoan cap carbonate sequence. Our results contrast with suggestions by previous workers that the Shuram excursion followed the Gaskiers event by some 20–30 m.y. We suggest instead that the Shuram and Gaskiers events were contemporaneous with the biotrigraphi graphic transition from acanthomorphic to leiospherid acritarchs, and with the first appearance of widespread macroscopic animal life, 38 m.y. prior to the Ediacaran-Cambrian boundary.

INTRODUCTION

Ediacaran strata record a critical period in Earth history (635–541 Ma), during which metazoan life first appeared (Knoll et al., 2004, 2006; Narbonne et al., 2012). They also record a significant rise in atmospheric and oceanic oxygen (Fike et al., 2006; Canfield et al., 2007; McFadden et al., 2008; Sahoo et al., 2012), which was a prerequisite to metabolic function in animals (Knoll and Carroll, 1999; Och and Shields-Zhou, 2012). Neoproterozoic oxygenation resulted in atmospheric oxygen levels generally interpreted as similar to those of the present day (Holland, 2006; Kump, 2008). Today, atmospheric oxygen levels are maintained by photosynthesis from land plants and marine organisms in roughly equal proportions (e.g., Field et al., 1998). It has therefore long been enigmatic that land plants are not preserved in rocks older than ca. 400 Ma, or ~150 m.y. later than the first appearance of animals. For that reason, it is widely presumed that the rise of animal life required sufficient oxygen production, from either marine photosynthesis, or perhaps some sort of “bootstrap” mechanism from animals themselves, to survive (e.g., Butterfield, 2008; Lennon et al., 2014). In any event, progress toward understanding the fundamental question, “what is the origin of animals?”, hinges in part on understanding how and when oxygen became sufficiently available to make animal metabolism possible (e.g., Nussal, 1959).

Among the most fruitful avenues of research along these lines to date has been exploration of proxies for the chemistry of seawater in which animal life first appeared, primarily the stable isotope geochemistry of shallow-marine carbonate strata. The best-preserved Ediacaran strata around the globe that contain carbonate all feature a singularly large (by about a factor of two) negative anomaly in the isotopic composition of carbon, which has been attributed primarily to the isotopic composition of ancient seawater itself (Fike et al., 2006; McFadden et al., 2008; but for an alternative view, see Swart and Kennedy, 2012). The anomaly is best preserved and documented in the Ediacaran Shuram Formation in Oman (Burns and Matter, 1993; Le Guerroué et al., 2006a, 2006b; Osburn et al., 2015), and it is generally referred to as the “Shuram excursion,” taking its name from the discovery formation. A similar excursion has been documented in Neoproterozoic sections on five of Earth’s seven modern continents, and it occurs only once in each section: Africa (Kaufman et al., 1991; Halverson et al., 2005), Asia (Burns and Matter, 1993; Condon et al., 2005; Melezhik et al., 2005; Fike et al., 2006; McFadden et al., 2008; Macdonald et al., 2009; Osburn et al., 2015), Australia (Calver, 2000; Husson et al., 2015), Europe (Melezhik et al., 2005; Prave et al., 2009), and North America (Myrow and Kaufman, 1999; Corsetti et al., 2000; Corsetti and Kaufman, 2003; Kaufman et al., 2007; Bergmann et al., 2011; Petterson et al., 2011; Verdel et al., 2011;
Macdonald et al., 2013). The Shuram excursion is the largest known Neo-
proterozoic or younger carbon isotope anomaly (Grotzinger et al., 2011), and
its magnitude is among the largest recorded in Earth history (see, for example,
the Paleoproterozoic Lomagundi-Jatuli excursion; Bekker and Holland, 2012).

Global chronostratigraphic expression of the Shuram excursion is a re-
markable discovery from at least three perspectives. First, it represents a pre-
sumably isochronous fingerprint of a specific interval of time from sections
with notoriously sparse age constraints. Second, it implies that a geologically
extreme event of uncertain origin occurred at the same time as the rise of
animals. Last, the singular magnitude of the excursion contributes to the goal
of creating a global composite time series of secular variations in marine car-
bon isotope ratios. In regard to the third point, the duration of the anomaly
raises the potential for using the shapes of the curves, rather than simply
the magnitudes of the excursions, as a correlation tool from section to sec-
tion; this of course presumes a relatively constant sedimentation rate at the
hundred-meter scale (Halverson et al., 2005; Saltzman and Thomas, 2012). For
the Shuram excursion, δ13C values rapidly descend with stratigraphic position
to <−11‰, followed by a recovery that is at first gradual and then moderate in
slope, with the change occurring near −4‰ (fig. 2 in Condon et al., 2005; fig. 3
in Prave et al., 2008; fig. 16 in Verdel et al., 2011; fig. 3 in Grotzinger et al., 2011;
fig. 13 in Macdonald et al., 2013; fig. 1A in Husson et al., 2015).

At present, the most significant impediment to understanding Ediacaran
biostratigraphy is the lack of internal age control in most sections around the
globe. The ages of the boundaries of the Ediacaran Period are well defined
radiometrically in multiple sections. The base is defined by the lithologically
distinctive post-Marinoan cap carbonate sequence, which is associated with a
−6‰ δ13C excursion in carbonate and is precisely dated at 635 Ma in Namibia
and China (Hoffmann et al., 2004; Condon et al., 2005). The top is defined by
the first appearance of the trace fossil Treptichnus pedum (541 Ma), which is
also associated with a −6‰ δ13C excursion in carbonate. Other than the first
appearance of large Ediacaran body fossils, which usually occurs rather high
in most sections relative to the Ediacaran-Cambrian boundary, the Shuram
excursion has emerged as the single most distinctive stratigraphic datum that is
globally recognized. However, its precise age is poorly constrained, precluding
any attempt to meaningfully subdivide some 94 m.y. of Ediacaran time,
and creating first-order uncertainties in the relative timing of major environ-
mental and biostratigraphic events (Xiao et al., 2016). A second major strati-
graphic feature, largely restricted to sections in the North Atlantic region, is
the Gaskiers glaciation (Myrow and Kaufman, 1999), which, in contrast to the
Shuram event, is precisely dated at 579 Ma (Bowring et al., 2003a, 2003b; Pu
et al., 2016). The mismatch between sections with glaciogenic rocks and pre-
cise radiometric ages on one hand, and the Shuram excursion in carbonate
strata on the other, has left it uncertain whether or not these two events are
correlative (Xiao et al., 2016). A 580 Ma age for the Shuram excursion provides
an obvious correlation between the two most conspicuous events in the Edia-
caran record (e.g., Xiao et al., 2004; Fike et al., 2006; Zhou et al., 2007; Halver-
son et al., 2005, 2010; Loyd et al., 2012; Schiffbauer et al., 2016). Alternatively,
the stratigraphic proximity of the Shuram excursion to the Precambrian-Cam-
brian boundary, and a 551 Ma ash bed near the apparent upper zero crossing
of the excursion in the Doushantuo Formation of China, suggest that it may be
as much as 20–30 m.y. younger than the Gaskiers glaciation (Condon et al.,
2005; Bowring et al., 2007; Cohen et al., 2009; Sawaki et al., 2010; Narbonne
et al., 2012; Macdonald et al., 2013; Tahata et al., 2013; Xiao et al., 2016).

One chronological tool that has heretofore only been sparingly applied to
Ediacaran strata is thermal subsidence analysis (e.g., Le Guerroué et al.,
2006b). It is well known that thermal subsidence associated with seafloor
spreading is a useful chronometer that can predict the age of the ocean floor
based on the exponential decay of its elevation with respect to the abyssal
plains for lithosphere older than 20 m.y. (e.g., equation 22 in Parsons and
Sclater, 1977). The same principle also applies to models of the subsidence
history of passive-margin basins, which include an initial thickness of newly
stretched continental crust and substantial sediment loading (McKenzie, 1978).

The decay is predicted by laws of diffusive heat transport of physical rigor
that are on par with laws of closed-system radioactive decay used to date the
timing of crystallization of minerals. The principal limitations in using thermal
conduction as a chronometer are (1) the requirement that subsidence records
thermal relaxation without significant mechanical modification of the litho-
sphere, such as extension, flexural loading, instability of a thermal boundary
layer, or unmodeled sources of dynamic topography; and (2) corrections of
the observed stratigraphic subsidence for the compaction and lithification
of sediment after deposition, and for water depth and changes in sea level
(e.g., Steckler and Watts, 1978; Allen and Allen, 2005).

In comparison with Phanerozoic sedimentary basins, published sub-
sidence analyses of Ediacaran strata have been limited, with most of the effort
thus far concentrated on the western Laurentian continental margin (Stewart
and Suczek, 1977; Bond et al., 1983; Armin and Mayer, 1983; Levy and Christie-
Blick, 1991; Yankee et al., 2014). The focus on this region as a testing ground
for thermal subsidence modeling was due to the fact that it is perhaps the
best-preserved example of an ancient passive margin, analogous to pres-
ent-day Atlantic-type margins, but with virtually complete surface exposure of
apparent synrift and postrift sedimentary archives spanning several hundred
million years (Stewart, 1972; Gabrielse, 1972; Burchfiel and Davis, 1972, 1975;
Stewart and Poole, 1974; Dickinson, 1977; Monger and Price, 1979). Because
these sequences span the Ediacaran-Cambrian boundary, such that roughly
half their thickness is Proterozoic in age, temporal control on subsidence has
been restricted mainly to the Phanerozoic portion of subsidence curves. The
lack of age control on the lower part of the section precludes precise defini-
tion of the transition from mechanical extension to pure thermal subsidence.
Fortunately, the accurate definition of an exponentially decaying system, in
particular, extrapolating stratigraphic age backward in time from a curve with
known ages, is independent of the timing of onset and total amount of purely
thermal subsidence.

Here, we address the problem of the correlation and age of the Shuram
isotopic excursion through lithostratigraphic and chronostratigraphic study of
the type locality of the Ediacaran Johnnie Formation in the Spring Mountains of southern Nevada. The Johnnie Formation is at least 1800 m thick at the type locality, and it makes up more than half of the maximum known thickness of ~3000 m of total Ediacaran strata exposed in this region. The underlying Noonday Formation provided the first isotopic match between the Marinoan cap carbonate sequence in Namibia (Hoffman et al., 1998) and a section from another continent (Petterson et al., 2011). The overlying Stirling and Wood Canyon Formations contain Ediacaran and Lower Cambrian fossil assemblages that define the Cambrian-Precambrian boundary within the lower part of the Wood Canyon Formation (Corsetti and Hagadorn, 2000; Hagadorn and Wagganer, 2000), 1200 m above the top of the Johnnie Formation in the Spring Mountains. The uppermost 300 m of section of the Johnnie Formation contains the best expression of the Shuram excursion in Laurentia (Corsetti and Kaufman, 2003; Kaufman et al., 2007; Bergmann et al., 2011; Verdel et al., 2011). Therefore, to the extent that the section was deposited at or very near sea level on a thermally subsiding continental shelf, subsidence analysis may be used to estimate the age of the Shuram excursion and perhaps even broadly constrain the overall age of the Johnnie Formation.

### GEOLOGIC SETTING

Neoproterozoic-Cambrian strata in western Laurentia are divisible into two principal components, including a lower diamictite and volcanic sequence, and an upper terrigenous detrital sequence (Stewart and Suczek, 1977; Poole et al., 1992). The Johnnie Formation is the lowest siliciclastic formation in the upper terrigenous detrital sequence, forming the basal deposits of a westward-thickening continental margin terrace wedge, widely regarded to have developed in the wake of late Neoproterozoic rifting of the Rodinian supercontinent (Li et al., 2008, 2013). The formation is a few hundred meters thick near its eastern pinchout beneath Lower Cambrian cratonic strata, systematically increasing to at least 1500 m thick in its westernmost exposures, where the base is not definitively exposed (Stewart, 1970; this report). Lithologically, it is primarily variegated siltstone and very fine-grained sandstone that contains varying amounts (10%–40%) of carbonate and orthoquartzite, distinguishing it from the carbonate-dominated Noonday Formation below and coarse siliciclastic rocks of the Stirling Formation above (Fig. 1).

The Johnnie Formation was first defined and described in the northwest Spring Mountains in the Johnnie Wash area (Fig. 2; Nolan, 1924, 1929), where its contact with the underlying Noonday Formation is apparently not exposed, and hence its thickness is a minimum for this location. Nolan’s (1924) thickness and description were included in the regional stratigraphic synthesis of Stewart (1970). The type locality was subsequently mapped and briefly described by Burchfiel (1964, 1965), and relatively complete lithostratigraphic sections were measured by Hamill (1966) and Benmore (1978). The type locality has since received little attention in comparison to the much thinner sections in the Nopah Range and environs 70 km to the south, or equivalents 100 km to the west in the Panamint Range, where its basal contact with the Noonday Formation is extensively exposed (e.g., Hazzard, 1937; Wright and Troxel, 1966; Labotka et al., 1980; Albee et al., 1981; Benmore, 1978; Summa, 1993; Fedo and Cooper, 2001; Corsetti and Kaufman, 2003; Kaufman et al., 2007; Verdel et al., 2011). With the exceptions of detailed studies of parts of the formation (Summa, 1993; Abolins, 1999; Bergmann et al., 2011), no systematic attempt has yet been made to describe and interpret the entire formation at its type locality in terms of key bed forms, depositional environments, sequence architecture, or chronostratigraphy, at the level of more southerly or westerly sections.

The uppermost part of the Johnnie Formation, the Rainstorm Member, is a lithostratigraphically distinctive unit that can be correlated with confidence over a broad region of southwestern North America, including eastern California and southern Nevada (Stewart, 1970), and it probably occurs as far south as northern Sonora, Mexico, where it forms a part of the Clemente Formation (Stewart et al., 1984). The basal strata of the Rainstorm Member are its most distinctive part. They include a thin (~2 m), siltstone-enveloped, regionally extensive oolitic marker bed known as the “Johnnie oolite” (e.g., Bergmann et al., 2011). The oolite is underlain by greenish gray siltstone, and it is over-
laid by distinctive pale-red, fine-grained sandstone with or without associated sandy or silty micrites (“liver-colored limestones”). The overlying units characteristically contain groove marks, flute casts, intraformational conglomerate, and other indicators of shallow-water, high-energy currents. These carbonates record the onset and most negative part of the Shuram excursion in eastern California and southern Nevada (Corsetti and Kaufman, 2003; Kaufman et al., 2007; Verdel et al., 2011), as well as in the Sonora sections (Loyd et al., 2012). Similar to formation-scale thickness variations in the terrigenous detrital sequence as a whole, the Rainstorm Member generally thickens westward from as little as 20 m in the thinnest measured section to more than 300 m in the thickest sections (Stewart, 1970; Verdel et al., 2011).

Lower and middle Johnnie Formation strata are sufficiently variable in their lithostratigraphy that recognition of regionally mappable members is not as straightforward as in the case of the Rainstorm Member. As noted by Summa (1993), sub–Rainstorm Member depositional settings of the Johnnie Formation are interpreted as inner-shelf to tidally influenced nearshore environments that were highly susceptible to sea-level fluctuation (Benmore, 1978; Fedo and Cooper, 2001; Schoenborn et al., 2012). As we describe herein, depositional environments tend to be more landward to the south and east in these units, as suggested by the abundance versus absence of desiccation features, fluvial versus marine deposition, and medium- to coarse-grained sandstones versus fine- to medium-grained sandstones. Although this variability complicates simple lithostratigraphic correlation, if interpreted correctly, it can be used as an effective indicator of sea-level rise and fall.

Reported age constraints from the Johnnie and correlative Clemente formations include (1) a 640 Ma U-Pb age from a single detrital zircon grain in sub–Rainstorm Member siltstones in the Panamint Range of eastern California (Verdel et al., 2011), and (2) potential Ediacaran body and trace fossils (e.g., Cyclomedusa plana and Palaeophycus tubularis, respectively) ~75 m below the oolite in the Clemente Formation (McMenamin, 1996). The U-Pb age, because it is based on a single grain, is subject to the uncertainty of contamination during mineral processing and needs to be confirmed with duplicate analyses. The putative fossils have been questioned after examination by other paleontologists (e.g., J.W. Hagadorn, 2017, personal commun.), and they have generally not been accepted in subsequent stratigraphic studies of the region (e.g., Loyd et al., 2012, 2015). Latest Ediacaran fossils have been recovered from the uppermost Stirling Formation and the Lower Member of the Wood Canyon Formation in the Spring Mountains and neighboring Montgomer Mountains to the south (e.g., Cloudina and Swartpuntia; Hagadorn and Waggner, 2000; Smith et al., 2017), from sections in stratigraphic continuity with the type Johnnie Formation. These are succeeded immediately upward by Lower Cambrian trace fossils (Treptichnus pedum), which places the Ediacaran-Cambrian boundary in the Lower Member of the Wood Canyon Formation (Fig. 1; Corsetti and Hagadorn, 2000).

The underlying Noonday Formation has been interpreted as the cap carbonate sequence of the Marinoan “snowball Earth” glaciation (Pettersson et al., 2011), which by definition would place its base at the beginning of the Ediacaran-Cambrian boundary.
The Johnnie Formation’s basal contact with the Noonday Formation is lithostratigraphically gradational, transitioning from sandy dolostones of the upper Noonday Formation (Mahogany Flats Member of Petterson et al., 2011) to interstratified dolomitic sandstone and orthoquartzite in the lower Johnnie Formation (Transitional Member of Stewart, 1970). Although traditionally regarded as a conformable contact on the basis of this gradation (Hazzard, 1937; Stewart, 1970; Wright and Troxel, 1984), the identification of local karstic surfaces along the contact raises the possibility that it is a disconformity with a substantial depositional hiatus (Summa, 1993).

In terms of chemostratigraphic constraints on age, the conspicuous excursions to ~6% at the base and top of the Ediacaran section are well expressed in the southern Laurentian sections (e.g., Petterson et al., 2011; Smith et al., 2016). The presence of the Shuram excursion in the Rainstorm Member, despite its value as a correlation tool, does little to constrain the depositional age, because unlike the tightly constrained boundary excursions, hard chronological constraints are lacking, as noted already.

For almost a century, the terrigenous detrital sequence has been studied extensively on many different levels. Much of the early work focused on stratigraphic group-level packages that record the transition from Precambrian to Cambrian time (Nolan, 1929; Burchfiel, 1964; Stewart, 1970). More recent work on the Johnnie Formation has focused largely on outcrops in eastern California (Summa, 1993; Fedo and Cooper, 2001; Verdel et al., 2011; Schoenborn and Fedo, 2011; Schoenborn et al., 2012), or on specific features related to the Rainstorm Member, such as an incision-related conglomeratic member (Summa, 1993; Abolins, 1999; Abolins et al., 2000; Clapham and Corsetti, 2005; Verdel et al., 2011), giant ooids (Trower and Grotzinger, 2010), or detailed chemostratigraphy of the Johnnie oolite (Bergmann et al., 2011). The lower and middle portions of the Johnnie Formation have not been given as much detailed attention, except in areas close to the craton miogeoclinal hinge, where the Johnnie Formation is only a few hundred meters thick. The 1600 m stratigraphic thickness of sub–Shuram excursion Johnnie Formation at the type locality exceeds the thickness of any globally correlative Ediacaran strata of which we are aware. Furthermore, total Ediacaran stratigraphic thickness in southwestern Laurentia measures over 3000 m, greater than the approximate thicknesses of sections in Australia (2500 m), Oman (1500 m), and China (300 m). Strata of the lower and middle Johnnie Formation at its type locality therefore represent one of the best opportunities among sections globally to provide a relatively complete record of Ediacaran time prior to the Shuram excursion. An important gap in our understanding of Ediacaran chemostratigraphy is the paucity of carbonate strata below the Shuram anomaly in most sections. Of the major global sections that contain it, only the Oman example contains abundant carbonate in immediately underlying strata, the Khufai Formation. Discovery of corroborative carbonate-bearing strata in one or more sections around the globe would thus represent a significant step in expanding the global inventory of chemostratigraphic time series for a critical interval in Neoproterozoic time.

### METHODS

#### Lithostratigraphy

To identify a structurally intact section of the Johnnie Formation, we performed geologic mapping at 1:10,000 scale in the northwest Spring Mountains, Nevada, both of the type locality at Johnnie Wash, and in an area ~4 km to the southwest near Nevada Highway 160, 3 km west-southwest of Mount Schader (Fig. 2). We used the Mount Schader, Nevada 1:24,000 quadrangle map (U.S. Geological Survey, 1968) as a topographic base. Our field mapping spanned 9 d total between 21 April 2015 and 2 May 2015. The geologic maps were used to identify optimum transects for measuring stratigraphic section. The Mount Schader section was measured and sampled in detail using a Jacobs staff mounted with a Brunton compass set to the dip of bedding. For each stratigraphic subunit, we recorded: (1) fresh and weathered color of lithology using a Munsell color chart; (2) grain size; and (3) bedding thickness (supplemental text). Section was measured to the resolution of ~0.5 m (or finer in some instances, if warranted). The Johnnie Wash section was measured using geologic cross sections, and the general lithologic characteristics were recorded in the field during geologic mapping (see Appendix for unit names and descriptions).

#### Chemostratigraphy

For carbon and oxygen isotope chemostratigraphy, we collected samples at 0.3–1 m resolution in carbonate units. Samples from the upper ~400 m of sub–Rainstorm Member lithostratigraphic units were collected from the Mount Schader section during stratigraphic logging. Samples from two prominent carbonate horizons that occur below the deepest exposed strata of the Mount Schader section were collected in the Johnnie Wash locality of the Spring Mountains, and at a location ~3 km north of Johnnie Wash (locality A in Fig. 2A). In total, 107 centimeter-scale sample chips were collected for carbon and oxygen isotopic analysis, including 36 from the Johnnie Wash section and locality A and 71 from the Mount Schader section. In the laboratory, sample chips were sliced open using a diamond-bladed wet saw to expose fresh, unweathered surfaces. From the fresh surfaces, a high-speed rotary tool with a diamond-tipped drill bit was used to powder the sample. We carefully extracted ~0.1 mg of analyte from each sample chip, taking care to avoid any visible alteration or veining. Sample powder was loaded into vials, the air was purged and replaced with helium gas, and then digested in phosphoric acid at 72 °C for at least 1 h to evolve sufficient CO₂ gas for analysis. Carbon and oxygen isotope ratios were measured at Caltech using a Delta V Plus isotope mass spectrometer (“gas bench”). Our values for δ¹³C and δ¹⁸O are reported relative to the Vienna PeeDee belemnite (VPDB) standard in permil notation. We used Caltech’s laboratory working standards, which were calibrated to NBS 18 and NBS 19 and have uncertainties of ±0.1‰. Standards were measured once for every nine samples to assess systematic error.

*Supplemental Items. Six figures (Figures S1–S6), two tables (Tables S1 and S2), and supplemental text. Please visit https://doi.org/10.1130/GES01678.S1 or access the full-text article on www.gsapubs.org to view the Supplemental Items.*
Subsidence Analysis

Our tectonic subsidence analysis is based on stratigraphic thicknesses compiled from multiple sources for the northwest Spring Mountains, Nevada. The inner-shell to fluvial-deltaic facies of virtually all units within the terrigenous detrital sequence in this region suggest shallow-water deposition, removing the need for paleobathymetric correction (Levy and Christie-Blick, 1991). We used thicknesses from this study combined with thicknesses for overlying formations principally based on Stewart (1970) and Burchfiel et al. (1974). Our analysis encompasses known time points ranging from the Ediacaran-Cambrian boundary in the Lower Member of the Wood Canyon Formation (Corsetti and Hagadorn, 2000; Hagadorn and Waggoner, 2000; Smith et al., 2016, 2017) at 541 Ma, up through the Devonian-Mississippian boundary at the top of the Devils Gate Formation (Burchfiel et al., 1974) at 359 Ma (Ogg et al., 2016). The only previous attempt at a geohistory analysis of the Spring Mountains (Levy and Christie-Blick, 1991) was temporally constrained mainly by the Lower-Middle and Middle-Upper Cambrian boundaries, which were then deemed to be ~30 m.y. older than their currently accepted ages. We followed methods described in Allen and Allen (2005) to delithify and progressively unload (backstrip) the stratigraphic column in order to obtain the tectonic component of subsidence. Delithification parameters for siliciclastic rocks were taken from table 9.1 in Allen and Allen (2005), and parameters for carbonate rocks were taken from Equation 3 in Halley and Schmoker (1983). Tectonic subsidence curves were calculated using Backstrip, an open-source software for decompaction and tectonic subsidence calculations (Cardozo, 2009). Results for our earliest model runs were verified by hand using a spreadsheet program (e.g., Larrieu, 1995).

RESULTS

Lithostratigraphy

The most salient feature of the Johnnie Formation in the Johnnie Wash type locality (Fig. 3) is that, although very fine-grained sandstone and siltstone are present in all mappable units (distinguishing it from the overlying Stirling and underlying Noonday formations), three intervals are characterized by an abundance of fine- to medium-grained sandstone (identified with Roman numerals I, II, and III on Fig. 4). The sand-rich intervals range from 160 to 430 m thick, form distinct, resistant ridges within the otherwise recessive Johnnie Formation, and establish a basis for subdividing it into mappable units. Each sand-rich interval exhibits characteristics that readily distinguish it from the other two, in terms of either bed forms (intervals I and II) or parasequence architecture (interval III). Further subdivision of the formation is afforded by a conspicuous, 30–40-m-thick cherty carbonate unit near the middle of the section, and by lithological variation within sand-rich interval III. Our subdivision consists of mappable units A, C, and E. The cherty carbonate marker and overlying sand-poor strata define units F and G. The upper sand-rich interval exhibits rhythmic variations of sandstone, siltstone, and carbonate that are divided into five units, H through L, each of which is defined at the base of a 30–100-m-thick sand-rich subinterval (Figs. 3 and 4; Fig. S1 [footnote 1]).

At the Johnnie Wash type section, bedding strikes approximately north-south and dips moderately to steeply eastward (Fig. 3). The total thickness of sub-Rainstorm Member strata is 1595 m. The lowest stratigraphic unit (unit A) encountered is a recessive, slope-forming phyllitic siltstone that contains a continuous cleavage at high angle to bedding. The base of unit B is defined by the lowest occurrence of meter-scale orthoquartzite beds, which are abundant in the unit. Unit B is readily distinguished from higher sand-rich intervals by pervasive penecontemporaneous deformation. Nearly every sandstone horizon is affected, principally by ball-and-pillow structure, so much so that individual orthoquartzite beds are difficult to trace along strike for more than a few tens of meters. Individual ball-and-pillow structures are up to meter-scale in size (roughly equal to orthoquartzite bed thickness) and occur where fine-grained sandstone and siltstone underlie coarser sandstone. The ball-and-pillow structure is manifested in some places as simple load casts with folded lamination (Fig. 5A) and in others as completely detached sand bodies that have slumped downward into the underlying siltstone, surrounded by flame structure developed within the siltstone (Fig. 5B). At the type locality, the occurrence of ball-and-pillow structure in the Johnnie Formation is restricted to unit B, but it was also observed along one horizon at the top of the Rainstorm Member in the Mount Schader section (Figs. 6 and 7).

Unit C marks a return to generally incontinuous, slope-forming siltstone with a prominent orthoquartzite marker horizon near the middle of the unit. The uppermost beds include a brown, resistant, 2-m-thick dolostone bed, which marks the lowest occurrence of carbonate in the type section.

Unit D includes orthoquartzite and less abundant siltstone. The sedimentary characteristics that distinguishes unit D from the other two sand-rich intervals is abundant high-angle cross-stratification, preserved in medium- to thick-bedded orthoquartzite (Fig. 8A). Millimeter- to centimeter-scale laminae or thin beds are preserved in foresets within decimeter- to meter-scale beds that can be followed for tens of meters along strike. Foreset lamination is consistently truncated at high angles, ranging from 20° to 30°, by overlying beds (Fig. 8B). In stratigraphic coordinates (corrected by tilting to horizontal about the line of strike), poles to foreset lamination are strongly unimodal, dispersed in trend by more than 90° around a mean vector of ~N30°E 65° (Fig. 8B). We measured grain-size variation with stratigraphic height across a sequence of about 10 foreset layers (Figs. 8C and 8D) to test for the presence of reverse grading, which is characteristic of dry grain flows on the lee side of dunes (e.g., Hunter, 1977; Boggs, 2012). The results indicated that the mean grain size of ~200 μm varies little through the sample, if anything fining slightly upward. In general, the lamination is not defined by concentrations of detrital heavy minerals. Opaque phases in these quartzites are largely diagenetic and relatively uniformly distributed throughout the rock.
Figure 3. Geologic map and cross sections of Johnnie Wash and environs. Definitions of unit labels and description of map units are given in the Appendix.
The boundary between units D and E is among the most readily mappable in the area and is also associated with a color change on remote imagery from dark brown to light brown, which is the most conspicuous color contrast in the section (Fig. 2A). Unit E is ~280 m thick, and it is dominated by very fine-grained sandstone and siltstone, generally lacking the mature, fine- to medium-grained quartzitic sandstone characteristic of unit D. Near the top, unit E contains an interval of ~30 medium-bedded cycles that alternate between massive, immature fine-grained sandstone and laminated siltstone. Unit F is a conspicuous, 30–40-m-thick, gray cherty dolostone (Fig. 5C) that can be followed for at least 5 km along strike, albeit with some minor faulting. Unit G is 135 m thick, and it returns to siltstone and very fine-grained sandstone similar to unit E, with a few inconspicuous orthoquartzite beds.

The overall lithostratigraphic character changes beginning at the base of unit H, from relatively thick, homogeneous sandstone-, siltstone-, or carbonate-dominated units below to the far more compositionally heterogeneous units above. From the base of unit H up to the base of the Rainstorm Member, the section contains abundant orthoquartzite, defining the uppermost of the three sand-rich intervals in the Johnnie Formation (Fig. 4). For mapping purposes, the most straightforward subdivision of sand-rich interval III in the Johnnie Wash area is defined by five quartzite-dominated subunits ranging from 10 to 100 m thick, which define the lower parts of units H, I, J, K, and L (Fig. 4; Fig. S1 [footnote 1]). Each of these subunits is overlain by variable thicknesses of recessive, variegated siltstone (Fig. 5E). The occurrence of carbonate is sporadic. In the section in Johnnie Wash, units I, J, and K are each capped by a resistant, 1–3-m-thick subunit of brown-weathering, laminated dolostone (Fig. 5D), and units H and L do not contain carbonate (Fig. 4). The Mount Schader section contains a lesser proportion of quartzitic sandstone and a greater proportion of carbonate and siltstone (Figs. 6 and 7), which is the basis for selecting it, instead of the type locality, for detailed measurement and chemostatigraphic sampling. Even within the Johnnie Wash area, the distribution of quartzite, siltstone, and carbonate changes along strike, on a scale of a few kilometers (Abolins, 1999). Although orthoquartzite beds in units H through L locally exhibit some high-angle cross-stratification in the Johnnie Wash section, they contrast with unit D (sand-rich interval II) in mainly being parallel-bedded or, in the case of the Mount Schader section, hummocky cross-stratified. Orthoquartzite in units H–L is generally fine- to medium-grained and appears to contrast with lower sand-rich intervals in containing a greater proportion of medium-grained and locally coarse-grained sand.

Our informal unit nomenclature ends at the base of the formally defined Rainstorm Member, which, as noted earlier, is readily identified throughout the region on the basis of lithologic characteristics. In the northern Spring Mountains, the Rainstorm Member contrasts with the underlying units H through L in lacking fine- to medium-grained orthoquartzite beds. At the base of the Rainstorm Member, a fissile, phylitic siltstone is overlain by the ochre-colored, 2-m-thick Johnnie oolite. The ooids are up to ~2 mm in diameter (Fig. 5F) and exhibit local cross-stratification. The oolite horizon has erosional basal and upper contacts, locally including intraformational breccia and conglomerate, which includes cobbles and small boulders of the oolite. Above the Johnnie oolite, pale-red limestones locally contain dispersed, coarse quartz grains interstratified with carbonate-cemented, fine-grained sandstone. These carbonate-rich rocks are overlain by argillaceous mudstone with interbedded limestones, with scattered horizons of intraformational conglomerate.

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| Roman Numerals | Lithologic Description |
|---------------|------------------------|
| I             | Phylitic siltstone with ball-and-pillow structure |
| II            | Cyclic sandstone & laminated siltstone |
| III           | Silstone & occasional orthoquartzite |
|               | Granular orthoquartzite & siltstone, with dolostone marker bed near top |
|               | Orthoquartzite & siltstone |
|               | Orthoquartzite, siltstone, & dolostone |
|               | Orthoquartzite & siltstone |
|               | Orthoquartzite, siltstone, & dolostone |
|               | Orthoquartzite, sandstone near top |
|               | Orthoquartzite & siltstone |
|               | A | Phylitic siltstone (Paddys fault) |
|               | B | Phylitic siltstone & orthoquartzite with ball-and-pillow structure |
|               | C | Siltstone with orthoquartzite marker bed near middle & carbonate near top |
|               | D | Orthoquartzite with high-angle cross-bedding & siltstone |
|               | E | Cyclic sandstone & laminated siltstone |
|               | F | Orthoquartzite & siltstone |
|               | G | Siltstone & occasional orthoquartzite |
|               | H | Orthoquartzite, siltstone, & dolostone |
|               | I | Granular orthoquartzite & siltstone, with dolostone marker bed near top |
|               | J | Orthoquartzite & siltstone |
|               | K | Orthoquartzite & siltstone |
|               | L | Orthoquartzite & siltstone |
|               | M | Orthoquartzite & siltstone |
|               | N | Orthoquartzite & siltstone |
|               | O | Orthoquartzite & siltstone |
|               | P | Orthoquartzite & siltstone |
|               | Q | Orthoquartzite & siltstone |
|               | R | Orthoquartzite & siltstone |
|               | S | Orthoquartzite & siltstone |
|               | T | Orthoquartzite & siltstone |
|               | U | Orthoquartzite & siltstone |
|               | V | Orthoquartzite & siltstone |
|               | W | Orthoquartzite & siltstone |
|               | X | Orthoquartzite & siltstone |
|               | Y | Orthoquartzite & siltstone |
|               | Z | Orthoquartzite & siltstone |

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**Figure 4.** Generalized lithostratigraphic column of the Johnnie Formation at its type locality in Johnnie Wash, with thicknesses based on cross sections A-A’ and B-B’ in Figure 3 for units A through L, and Roman numerals on left side of column indicate the three sand-rich intervals discussed in text.
Figure 5. Photographs of selected lithostratigraphic elements of the Johnnie Formation. 

(A) Load cast with folded laminae in sandstone bed, unit B, Johnnie Wash area. Hammer is 28 cm long. 

(B) Ball-and-pillow structure in unit B, with light-colored siltstone (just above pocket knife) protruding upward between bulbous masses of fine-grained sandstone. Pocket knife is 9 cm long. 

(C) Gray cherty dolostone, unit F, Johnnie Wash. Pencil is 15 cm long. 

(D) Brown-weathering, laminated dolostone, typical of carbonate beds in units H through L. Hammer is 28 cm long. 

(E) Variegated siltstone typical of all units in the Johnnie Formation. Pocket knife is 9 cm long. 

(F) Johnnie oolite, Johnnie Wash, ooids are 1–2 mm in diameter. 

(G) Large, bulbous mass of orthoquartzite (left of hammer), surrounded by smaller masses lying in a matrix of siltstone, uppermost bed of the Rainstorm Member, Mount Schader section. Hammer is 33 cm long. 

(H) Base of orthoquartzite bed in G, showing load cast structures with underlying siltstone. Hammer is 33 cm long.
The uppermost part of the Rainstorm Member in the Mount Schader section contains a 4-m-thick triad of quartzite, siltstone, and dolostone, in which the quartzite is disrupted by locally intense ball-and-pillow structure (Figs. 5G and 5H). The base of the overlying A Member of the Stirling Formation is marked by highly resistant, massively textured to cross-stratified, medium- to thick-bedded, medium- to coarse-grained orthoquartzite. The principal contrast between the Stirling Formation’s A Member and any of the orthoquartzites in the Johnnie Formation is the coarse grain size, including the common occurrence of granules and small pebbles of vein quartz and jasper. Neither the Johnnie Wash section nor the Mount Schader section appears to preserve the incised valley fill characteristic of the conglomeratic member of the Johnnie Formation (Abolins, 1999; Verdel et al., 2011).

**Chem stratigraphy**

Carbon isotope ratios range from a low of −4.4‰ (VPDB) to a high of 4.9‰ (Table S1 [footnote 1]). The data are mainly concentrated in carbonate beds in units H through L, which constitute the uppermost 400 m of pre-Rainstorm Member strata (Fig. 7). Within the underlying ~1100 m of exposed Johnnie Formation strata, carbonate intervals are present only in units C and F, 1440 m and 860 m below the base of the Rainstorm Member, respectively (Figs. 4 and 9). The lowest and highest values of δ<sup>13</sup>C occur in the stratigraphically highest samples, and they define a strong negative trend, beginning near the top of unit K and ending at the top of the Johnnie oolite bed. Below this marked trend in the data, there is otherwise no general trend.
Figure 7. (A) Detailed lithostratigraphic column of the Mount Schader section, from unit G of the Johnnie Formation through the lowermost part of the A Member of the Stirling Formation, showing thicknesses based on Jacobs staff measurements (transsects shown in Fig. 6). Detailed descriptions of subunits 1-48 are given in the supplemental text (text footnote 1). Detailed descriptions of subunits 1-48 are given in the supplemental text (text footnote 1). VPDB—Vienna Peedee belemnite.

Stirling Formation: orthoquartzite (dolomitic, cross-stratified, massive, or brecciated)

Rainstorm Member: siltstone, Johnnie oolite, sandy limestone, orthoquartzite with ball-and-pillow structure

siltstone with rare orthoquartzite, & brown dolostone

ripple-bedded siltstone & brown dolostone

siltstone, hummocky cross-stratified orthoquartzite, & brown dolostone

hummocky cross-stratified orthoquartzite, siltstone, & brown dolostone

siltstone, orthoquartzite & brown dolostone

siltstone & occasional orthoquartzite
More than 90% of the values recorded from the bottom of unit K to the unit C carbonate are positive, averaging 1.5‰, with a standard deviation of 1.2‰. The scatter in values within individual carbonate intervals is approximately the same as variations in the average values between carbonate intervals (Fig. 7). However, the variation in δ13C values with stratigraphic position within each relatively thin carbonate interval does not appear to be entirely random (Fig. 10). For example, carbonates from units C and the lower part of unit K show decreasing δ13C values stratigraphically upward (R² = 0.74, 0.88 respectively), whereas values from the lower two carbonates in unit H and the upper carbonate interval in unit J suggest increasing δ13C values stratigraphically upward (R² = 0.26–0.83).

Oxygen isotope ratios range from a low of –16.0‰ (VPDB) to a high of –5.0‰, with an average value of –9.4‰ (Table S1 [footnote 1]). There is no general trend in the mean values for each individual carbonate interval with stratigraphic position (Fig. 10C). The range of values within the carbonate intervals is as great as 6‰, i.e., greater than the variation of mean values for each interval. Correlation of δ18O and δ13C is poor for the data set as a whole (Fig. 11). Plots of δ18O versus stratigraphic position and side-by-side comparison with δ13C values

Figure 8. (A) Photograph of tabular planar cross-stratification in unit D, Johnnie Wash. Notebook is 13 cm wide. (B) Equal-area stereogram of poles to foreset lamina tions measured at locality A (Fig. 2A), with bedding tilt removed. Dots are data (n = 65); larger square in circle is the mean vector. Plotted using Stereonet software (Allmendinger et al., 2012; Cardozo and Allmendinger, 2013), with Kamb contours at 2σ intervals (Kamb, 1959). (C) Photograph of steep foreset laminations (bedding parallel to base of photograph), showing 9 cm scale bar at location of petrographic measurements of mean grain size in D. (D) Mean grain size as a function of stratigraphic height in sample of foreset laminations, showing relatively constant value of 200 µm.
are presented in the Supplemental Items (Fig. S2 [footnote 1]). Correlation of $\delta^{18}$O values with stratigraphic position within each interval is also generally poor. Of 12 beds with more than three samples, $R^2 > 0.5$ only for beds Zj1, Zj2, and Zj3 (see Table 1 for nomenclature). With regard to correlation of $\delta^{18}$O with $\delta^{13}$C, only the carbonate in unit C shows good positive correlation ($R^2 = 0.9$), but this interval only has four data points. Intervals with 10 or more data points all show poor ($R^2 < 0.1$) intrabed correlation of $\delta^{18}$O with $\delta^{13}$C (Fig. S2 [footnote 1]).

**Subsidence Analysis**

Our analyses focused on modeling the tectonic component of subsidence for strata in the Spring Mountains section: Johnnie unit A through the Devonian Devils Gate Formation (Table 1). The model results define the relationship between the stratigraphic thickness $S$ and the tectonic component of subsidence $Y$ (Table 2), which yields a resulting curve for the function $Y(S)$ (Fig. 12). This curve depends on parameters that describe lithification and isostatic adjustment due to sediment loading (Tables 2 and 3), and it is independent of time (Eqs. 1 and 2 in Steckler and Watts, 1978). We model the time dependence of subsidence in the Discussion section below.

Our determinations of $Y(S)$ include the effects of some 3000 m of Mississippian through Triassic overburden that lay above the Johnnie–Devils Gate interval during Jurassic and Early Cretaceous time (Giallorenzo et al., 2017). They also include two major sources of uncertainty. The first is the possible effect of a significant sedimentary substrate, predating Johnnie unit A, on the calculated tectonic subsidence. The substrate may either have been: (1) limited to the Noonday Formation or its equivalents, which are at most a few hundred meters thick and may be represented by the lowest units of the Johnnie Wash section (values $Y_{\text{ns}}$ indicate “no substrate”); or (2) a thick succession of Proterozoic Pahrump Group strata (Crystal Spring through Kingston Peak Formations), which could be present at depth beneath the northwest Spring Mountains (values $Y_{\text{ws}}$ indicate “with substrate”); note that in Fig. 12, $Y_{\text{ws}}$ values were plotted using the base of the Spring Mountains section as a datum for zero, for a direct comparison to $Y_{\text{ns}}$. The oldest Pahrump Group strata, the Crystal Spring Formation, were ~500 m.y. old in Ediacaran time, and therefore these models may somewhat overestimate the effect of sedimentary substrate on late Cryogenian–Ediacaran subsidence. The second major source of uncertainty lies in the resulting density of the delithified sediment column (Bond and Kominz, 1984; Bond et al., 1988). We simulated this error by varying sediment grain density by ±5%, and we note that the effect of sediment unloading is such that the lowest assumed density results in the highest tectonic component of subsidence, and vice versa. This density range yields variations in values of $Y$ for a given $S$ ($Y_{\text{ns}}$ or $Y_{\text{ws}}$; Table 4) that are similar to those obtained by Bond and Kominz (1984) and Levy and Christie-Blick (1991).

The resulting plots for $Y(S)$ (Fig. 12) show a decreasing ratio of tectonic subsidence per meter of sediment thickness, with slopes ($\Delta Y/\Delta S$) ranging from values near 1.0 at the base of the section for the “no substrate” curves, to as little as 0.1 near the middle of the section for the “with substrate” curve. More typically, slopes range from 0.3 to 0.6. There is an abrupt change in slope at $S = 3500$ m, where the section transitions from predominantly siliciclastic to predominantly carbonate sedimentation. On the no-substrate curve, the slopes defined by the five values closest to $S = 3500$ m are 0.5 ($S < 3500$ m) and 0.2 ($S > 3500$ m), with each of the two arrays appearing quite linear. Corresponding values on the “with substrate” curve are 0.4 and 0.1. Thus, although there is a degree of gradual curvature above and below $S = 3500$ m, most of the flattening of $Y(S)$ is associated with the lithologic transition.

The effect of including a thick substrate of Pahrump Group strata is to greatly reduce our estimate of $Y$ for any given $S$. In other words, by not accounting for the substrate, we overestimate the tectonic component of subsidence by 50% or more, particularly in the early phases of subsidence. Physically, the reason for this is that the no-substrate model inadvertently places incompressible basement rocks where a compacting substrate exists in the event that there is substrate, and therefore the model incorrectly assigns the compaction of the substrate to tectonic subsidence, resulting in an overestimate.

![Figure 9. (A) Carbon and (B) oxygen isotope ratios in carbonate as a function of stratigraphic position for units C and F in the Johnnie Wash area. Numerical values are found in Table S1 (text footnote 1). VPDB—Vienna Peedee belemnite.](image-url)
Figure 10. Carbon isotope ratios as function of stratigraphic position, expanding the vertical scale within each carbonate bed to reveal any intrabed trends. Beds are numbered from bottom to top within a given unit, e.g., Zjj1 is the lowest carbonate bed in unit J. VPDB—Vienna Peedee belemnite.
The uncertainties in Y due to sediment grain density are generally in the ±10%–15% range. In addition, to the extent that a thick sedimentary substrate is present below the lowest exposures of the Johnnie Formation in Johnnie Wash, tectonic subsidence may be overestimated by several tens of percent. Despite the sensitivity of both the density and substrate effects on the absolute value of Y, as we will discuss in the next section, the effect on estimating the age of tectonic subsidence is not large, because these estimates depend mainly on relative, not absolute values of Y. Specifically, (1) errors arising from density and substrate are correlated, such that Y(S) retains its shape even though Y may vary significantly; and (2) the exponential equation describing the time dependence of Y is defined by ratios between values of Y, rather than their absolute magnitudes.

## DISCUSSION

Perhaps the most basic question in regard to the origin of the Johnnie Formation is whether the sub-oolite interval contains recognizable subunits that can be correlated across its region of exposures, and the extent to which the section contains major unconformities. These issues are best addressed through lithostratigraphic characteristics and comparisons between the Spring Mountains section and the two other major sections in the region, the Desert Range to the north and the Nopah Range to the south. A second important question is whether or not the sub-oolite (sub–Shuram excursion) interval is a chemostratigraphic correlative with the sub–Shuram excursion Khufai Formation in Oman. A third significant issue is whether continuous Johnnie Formation (and subsequent) deposition occurred through most or all of Ediacaran time, because this aspect is critical to dating the Johnnie Formation using thermal subsidence modeling. A time-dependent exponential thermal subsidence model applied to our decompacted and backstripped subsidence model, Y(S) (Fig. 12), implies continuous sedimentation along the southwest Laurentian passive margin through the whole of Ediacaran time (i.e., from basal Noonday to early Wood Canyon time, or 635–541 Ma). If such a model is correct, it provides an independent estimate of the age and duration of the Shuram excursion, and whether or not it occurred near the time of the Gaskiers glaciation.

### Lithostratigraphy

Although lithostratigraphic correlation of sub–Rainstorm Member Johnnie Formation units is not as straightforward as for the overlying intervals, neither is it particularly complex. The two thickest sections, which both lie in Nevada, the northern part of the Johnnie outcrop belt, include the northern Spring Mountains and Desert Range sections. Both sections are readily divisible into alternating sand-rich and siltstone/carbonate-rich intervals,
each on the order of 100 m to a few hundred meters thick (I, II, III in Fig. 13 on the northern Spring Mountains section). The three sand-rich intervals of the Johnnie Formation are succeeded by four additional sand-rich intervals that have long been recognized as regionally correlative units (IV–VII in Fig. 13 on the northern Spring Mountains section), of which the top three have paleontological age constraints. The two sections each contain three sand-rich intervals below the Rainstorm Member that have proportionate relative thicknesses. Further, the siltstone/carbonate-rich interval between sand-rich intervals II and III contains an ~40-m-thick cherty dolostone unit in both sections, strengthening correlation, as noted by Stewart (1970) and Benmore (1978). We correlate sand-rich interval I of the Spring Mountains (unit B) with

### TABLE 1. NOMENCLATURE OF STRATIGRAPHIC UNITS USED IN SUBSIDENCE ANALYSIS TABLES

| Stratigraphic unit: can be partitioned/combined formations (Fm) and/or members (Mbr) | Abbreviation |
|---|---|
| Spring Mountains section | |
| Carbonate overburden | MzPzco |
| Devils Gate Fm | Ddg |
| Nevada Fm | Dn |
| Laketown Fm (upper 50%) | Dl |
| Laketown Fm (lower 50%) | Sl |
| Ely Springs Fm | Oes |
| Eureka Fm | Oe |
| Pogonip Group (upper third) | Op2 |
| Pogonip Group (lower two thirds) | Op1 |
| Nopah Fm (upper third) | OCN2 |
| Nopah Fm (lower two thirds) | OCN1 |
| Dunderberg Fm | Ed |
| Bonanza King Fm (Banded Mountain Mbr, upper 36%) | Ebk2 |
| Bonanza King Fm (Mbrs: Papoose Lake & Banded Mountain, lower 64%) | Ebk1 |
| Carrara Fm (upper two thirds) | Ec2 |
| Carrara Fm (lower third) | Ec1 |
| Zabriskie Fm | Ez |
| Wood Canyon Fm (Ediacaran-Cambrian boundary to top) | E2wc2 |
| Wood Canyon Fm (to Ediacaran-Cambrian boundary) | E2wc1 |
| Stirling Fm (members A through E) | Zsa - Zse |
| Johnnie Fm (Rainstorm Mbr, oolite bed’s base to top of Mbr) | Zjr2 |
| Johnnie Fm (Rainstorm Mbr, base to oolite bed’s base) | Zjr1 |
| Johnnie Fm (members A through L) | Zja - Zjl |
| Pahrump Group substrate (hypothetical) | |
| Johnnie Fm (presumed equivalent to the Transitional Mbr of Stewart, 1970) | Zjt |
| Kingston Peak Fm (Mbr: South Park, sub-Mbr: Wildrose) | Zkpw |
| Kingston Peak Fm (Mbr: South Park, sub-Mbr: Thorkndike) | Zkpth |
| Kingston Peak Fm (Mbr: South Park, sub-Mbr: Mountain Girl) | Zkpmg |
| Kingston Peak Fm (Mbr: South Park, sub-Mbrs: Sourdough & Middle Park) | Zkpsmp |
| Kingston Peak Fm (Limkiln-Surprise Mbr) | Zkspl |
| Beck Springs Fm/Kingston Peak Fm (lower) | Zbs |
| Horse Thief Springs Fm | Zhs |
| Crystal Springs Fm (upper) | Ycs2 |
| Crystal Springs Fm (lower) | Ycs1 |

### TABLE 2. NOMENCLATURE FOR PARAMETERS USED IN DELITHIFICATION AND BACKSTRIPPING ANALYSIS†

| Parameter | Description |
|---|---|
| $\phi_s$ | Surface porosity (%) |
| $c$ | Porosity depth coefficient (km$^{-1}$) |
| $\rho_{sg}$ | Sediment grain density (kg m$^{-3}$) |
| $h$ | Stratigraphic thickness (m) |
| $S$ | Cumulative stratigraphic thickness (m) |
| $S^*$ | Delithified/decompacted thickness (m) |
| $Y$ | Tectonic subsidence (m) |

†Subscripts for $S$, $S^*$, and $Y$: ns—no Pahrump Group substrate; ws—with Pahrump Group substrate; low—low subsidence, using $+5\%\rho_{sg}$ high—high subsidence, using $-5\%\rho_{sg}$.
the Carbonate member in the Desert Range section, on the basis of stratigraphic position. We note, however, that the pervasive soft sediment deformation in unit B has not been reported from orthoquartzites in the Carbonate member, and that unit B does not contain carbonate. Unit A, which is predominantly siltstone, would therefore correlate with siltstones and oolitic limestones underneath the Carbonate member. The oolitic limestone unit at the base of the Desert Range section has been considered to be correlative with the Noonday Formation (Longwell et al., 1965; Gillett and Van Alstine, 1982), implying that unit A in the Spring Mountains may also be a Noonday correlative (Fig. 13).

The southern Nopah Range section is approximately half the thickness of the northern Spring Mountains and Desert Range sections, and it contains a number of subaerial erosion surfaces that thus far have not been observed in the thicker Nevada sections (Summa, 1993). Like the Nevada sections, however, it does contain three sub–Rainstorm Member sand-rich intervals, suggesting lithostratigraphic correlation (Fig. 13). Specifically, the lower part of the Transitional, Quartzite, and Upper carbonate-bearing members of Stewart (1970) would correspond to sand-rich intervals I, II, and III, respectively, in the Spring Mountains. The correlation is strengthened by: (1) the alternating orthoquartzite/Carbonate cycles evident in sand-rich interval III in both the northern Spring Mountains and southern Nopah Range sections; (2) the lack of carbonate and abundance of high-angle cross-stratification in interval II in all three sections (Quartzite member = upper part of Lower quartzite and siltstone member = unit D, Fig. 13); (3) the consistency of unimodal, south-southwest-directed paleoflow directions in pre–Rainstorm Member orthoquartzites in the Spring Mountains and Desert Range sections (Fig. 14); and (4) the lithological similarity between the lowest sand-rich intervals in the southern Nopah Range and Desert Range sections, both of which contain a mixed carbonate-siliciclastic assemblage. Mitigating against these lithostratigraphic correlations are the observations that: (1) the interval III correlative in the Desert Range lacks carbonate; (2) interval I in the Spring Mountains (unit B) also lacks carbonate; and (3) the proposed Noonday substrate of interval I is lithostratigraphically dissimilar in all three sections, ranging from pale-gray quartz-rich dolomite boundstone in the Nopah Range, to phyllitic siltstone in the Spring Mountains, to medium-gray oolitic limestone in the Desert Range. Regardless of the details of these correlations, the most important facets of the two sections in Nevada are the following: (1) The sub–Rainstorm Member sections are at least twice as thick as the Nopah Range section; and (2) evidence for subaerial erosion, such as grikes, paleosols, channel scour, and desiccation cracks, which is conspicuous in the Nopah Range section, appears to be lacking. Although significant depositional hiatuses within the Nevada sections cannot be ruled out, the overall lithostratigraphic uniformity or “monotony” of these sections (siltstone and fine- to medium-grained sandstone and orthoquartzite, with sporadic thin carbonate beds) is consistent with conformable sedimentation on a stably subsiding continental shelf (Stewart, 1970; Fedo and Cooper, 2001; Schoenborn et al., 2012).
### TABLE 3. PARAMETERS USED IN DELITHIFICATION AND BACKSTRIPPING ANALYSIS OF THE SPRING MOUNTAINS SECTION

| Unit          | Age§          | Lithology / uni03D5 | c  | \(\rho_{sg}\)  | h  | \(S_n\)  | \(S_m\)  |
|---------------|---------------|---------------------|----|---------------|----|-----------|-----------|
| MzPzo         | 243           | l/d                 | 43 | 0.58          | 2785 | 3000      | 9720      | 11745     |
| Ddg           | 359           | d                   | 43 | 0.58          | 2710 | 286       | 6720      | 8745      |
| Dn            | 383           | d                   | 43 | 0.58          | 2860 | 286       | 6434      | 8459      |
| DI            | 383           | d                   | 43 | 0.58          | 2860 | 71.5      | 6148      | 8173      |
| Sl            | 419           | d                   | 43 | 0.58          | 2860 | 71.5      | 6076.5    | 8101.5    |
| Oes           | 444           | d                   | 43 | 0.58          | 2860 | 95        | 6005      | 8030      |
| Oe            | 458           | s                   | 49 | 0.27          | 2650 | 71        | 5910      | 7935      |
| Op2           | –             | l                   | 43 | 0.58          | 2710 | 230       | 5839      | 7864      |
| Op1**         | 470           | l                   | 43 | 0.58          | 2710 | 460       | 5609      | 7634      |
| OEn2          | –             | d                   | 43 | 0.58          | 2860 | 95        | 5149      | 7174      |
| OEn1          | 485           | d                   | 43 | 0.58          | 2860 | 191       | 5054      | 7079      |
| Cd            | –             | sh                  | 63 | 0.51          | 2720 | 48        | 4863      | 6888      |
| Ebk2          | –             | d                   | 43 | 0.58          | 2860 | 197       | 4815      | 6840      |
| Ebk1          | 497           | d                   | 43 | 0.58          | 2860 | 637       | 4618      | 6643      |
| Ec2           | –             | sh                  | 63 | 0.51          | 2720 | 286       | 3981      | 6006      |
| Ec1           | 509           | sh                  | 63 | 0.51          | 2720 | 143       | 3695      | 5720      |
| Ec            | –             | s                   | 49 | 0.27          | 2650 | 24        | 3552      | 5577      |
| ZCwz2         | –             | s/slt               | 49 | 0.27          | 2650 | 523       | 3528      | 5553      |
| ZCwz1         | 541           | s/lt/slt            | 49 | 0.27          | 2650 | 144       | 3005      | 5030      |
| Zs             | –             | s                   | 49 | 0.27          | 2650 | 340       | 2861      | 4886      |
| Zsd††         | –             | d                   | 43 | 0.43          | 2755 | 10        | 2521      | 4546      |
| Zsc           | –             | silt/s              | 49 | 0.27          | 2650 | 190       | 2511      | 4536      |
| Zsb           | –             | s/slt               | 49 | 0.27          | 2650 | 90        | 2321      | 4346      |
| Zsa           | –             | s                   | 49 | 0.27          | 2650 | 369       | 2231      | 4256      |
| Zg†±†         | –             | silt/s/d            | 47 | 0.37          | 2695 | 250       | 1962      | 3887      |
| Zr†††         | –             | silt                | 49 | 0.27          | 2650 | 17        | 1612      | 3637      |
| Zj††          | –             | silt                | 49 | 0.27          | 2650 | 60        | 1595      | 3620      |
| Zj             | –             | silt                | 49 | 0.27          | 2650 | 95        | 1535      | 3560      |
| Zj††          | –             | silt                | 49 | 0.27          | 2650 | 190       | 1440      | 3465      |
| Zj             | –             | silt                | 49 | 0.27          | 2650 | 55        | 1250      | 3275      |
| Zj†††         | –             | silt                | 49 | 0.27          | 2650 | 60        | 1195      | 3220      |
| Zg††          | –             | silt                | 49 | 0.27          | 2650 | 135       | 1135      | 3160      |
| Zf†††         | –             | d                   | 43 | 0.58          | 2860 | 40        | 1000      | 3025      |
| Zf              | –             | s/silt              | 49 | 0.27          | 2650 | 280       | 960       | 2985      |
| Zfd††††       | –             | s                  | 49 | 0.27          | 2650 | 300       | 680       | 2705      |
| Zp2           | –             | silt                | 49 | 0.27          | 2650 | 45        | 380       | 2405      |
| Zc1±±±±       | –             | silt                | 49 | 0.27          | 2650 | 50        | 335       | 2360      |
| Zcα†±±±±      | –             | s                  | 49 | 0.27          | 2650 | 60        | 1195      | 3220      |
| Zjα           | –             | silt                | 49 | 0.27          | 2650 | 125       | 125       | 2150      |
| Zj††††        | –             | d/s                 | 46 | 0.43          | 2755 | 125       | –         | 2025      |
| Zkpw†††††     | 635           | ss                 | 49 | 0.27          | 2650 | 100       | –         | 1900      |
| Zkpth         | –             | d                   | 43 | 0.58          | 2860 | 100       | –         | 1800      |
| Zkpng         | –             | s                  | 49 | 0.27          | 2650 | 100       | –         | 1700      |
| Zkpsm         | –             | silt                | 49 | 0.27          | 2650 | 200       | –         | 1600      |
| Zkpals        | –             | s/cgl               | 49 | 0.27          | 2650 | 400       | –         | 1400      |
| Zbs            | –             | d                   | 43 | 0.58          | 2650 | 200       | –         | 1000      |
| Zhs±±±±       | <787          | s                  | 49 | 0.27          | 2650 | 200       | –         | 800       |
| Ycs2          | >1087         | d                   | 43 | 0.58          | 2860 | 200       | –         | 600       |
| Ycs±±±±±±     | –             | s                  | 49 | 0.27          | 2650 | 400       | –         | 400       |

1Values from table 9.1 in Allen and Allen (2005), Equation 3 in Halley and Schmoker (1983), Deer et al. (1992), or weighted averages for lithologic mixtures. Abbreviations for lithology are: s—sandstone; slt—siltstone; l—limestone; d—dolostone; sh—shale; cgl—conglomerate. See Table 2 for parameter definitions.

2Ages at top of unit.
3Lithologic ratio used is 50/50; average for \(\rho_{sg}\).
4Carbonate is mostly limestone (fig. 2 in Burchfiel et al., 1974).
5Dolostone is sandy (stratigraphic column for Spring Mountains, plate 2 in Stewart, 1970).
6Shuram excursion ends. Lithologic ratio is slt+s/l/d/uni00A0= 67/16.5/16.5 (table 3 in Stewart, 1970).
7Shuram excursion begins.
8Ball-and-pillow structure.
9Lithologic ratio is d/s/uni00A0= 50/50.
10Age from Petterson et al. (2011).
11Maximum age from Mahon et al. (2014).
12Minimum age from Heaman and Grotzinger (1992).
| Unit          | Age† (Ma) | $S_{ns}$ | $S^*_n$ | $Y_{ns,low}$ | $Y_{ns,high}$ | $S_{ws}$ | $S^*_w$ | $Y_{ws,low}$ | $Y_{ws,high}$ | $S_{as}$ | $S^*_a$ | $Y_{as,low}$ | $Y_{as,high}$ |
|---------------|-----------|----------|---------|--------------|--------------|----------|---------|--------------|--------------|----------|---------|--------------|--------------|
| MzPzco        | 243       | 9720     | 9714    | 2670         | 3193         | 3715     | 11745   | 3128         | 3768         | 4406     |
| Ddg           | 359       | 6720     | 7489    | 2412         | 2791         | 3169     | 8745    | 9548         | 2905         | 3401     |
| Dn            | 383       | 6434     | 7246    | 2380         | 2743         | 3105     | 8459    | 9310         | 2878         | 3358     |
| Di            | 393       | 6148     | 7002    | 2367         | 2713         | 3058     | 8173    | 9071         | 2871         | 3334     |
| Sl            | 419       | 6076.5   | 6940    | 2364         | 2706         | 3046     | 8101.5  | 9011         | 2869         | 3328     |
| Oes           | 444       | 6005     | 6881    | 2361         | 2699         | 3035     | 8030    | 8953         | 2868         | 3322     |
| Oe            | 458       | 5910     | 6799    | 2357         | 2689         | 3020     | 7935    | 8874         | 2866         | 3315     |
| Op2           | –         | 5839     | –       | –            | –            | –        | 7864    | –            | –            | –        |
| Op1           | 470       | 5609     | 6541    | 2320         | 2636         | 2950     | 7634    | 8622         | 2836         | 3269     |
| OCh2          | –         | 5149     | –       | –            | –            | –        | 7174    | –            | –            | –        |
| OCh1          | 485       | 5054     | 6070    | 2280         | 2564         | 2846     | 7079    | 8167         | 2811         | 3212     |
| Ed            | –         | 4863     | –       | –            | –            | –        | 6888    | –            | –            | –        |
| Cbk2          | –         | 4815     | –       | –            | –            | –        | 6840    | –            | –            | –        |
| Cbk1          | 497       | 4618     | 5694    | 2264         | 2522         | 2778     | 6643    | 7804         | 2809         | 3183     |
| C2            | –         | 3981     | –       | –            | –            | –        | 6006    | –            | –            | –        |
| C1            | 509       | 3695     | 4798    | 2125         | 2327         | 2528     | 5720    | 6494         | 2710         | 3029     |
| Cz            | –         | 3552     | –       | –            | –            | –        | 5577    | –            | –            | –        |
| C ZwC1        | 541       | 3005     | 4024    | 1844         | 2009         | 2173     | 5030    | 6220         | 2475         | 2757     |
| C ZwC2        | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |
| C4            | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |
| C5            | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |
| C6            | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |
| C7            | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |
| C8            | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |
| C9            | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |
| C10           | –         | 3528     | –       | –            | –            | –        | 5553    | –            | –            | –        |

†See Table 2 for parameter definitions.
§Ages are at top of unit.
The pervasive ball-and-pillow and other paleoliquefaction structures in sand-rich interval I (unit B) are most simply interpreted as reflecting a period of high sediment flux during early Johnnie Formation deposition. These structures may have significance for the timing of the transition from mechanical stretching of the lithosphere to purely thermal subsidence, because (1) rapid subsidence is characteristic of both the rift phase and early thermal subsidence phase of passive-margin formation (e.g., Sawyer et al., 1982), and (2) such structures could be evidence for seismic shaking (e.g., Sims, 2012). The observation that essentially the entire 160 m thickness of unit B is affected implies that, whatever its cause, it was persistent over a sustained period of time. The other significant observation is that with only one exception, paleo liquefaction structures do not appear anywhere else higher in the section, despite

The pervasive ball-and-pillow and other paleoliquefaction structures in sand-rich interval I (unit B) are most simply interpreted as reflecting a period of high sediment flux during early Johnnie Formation deposition. These structures may have significance for the timing of the transition from mechanical stretching of the lithosphere to purely thermal subsidence, because (1) rapid subsidence is characteristic of both the rift phase and early thermal subsidence phase of passive-margin formation (e.g., Sawyer et al., 1982), and (2) such structures could be evidence for seismic shaking (e.g., Sims, 2012). The observation that essentially the entire 160 m thickness of unit B is affected implies that, whatever its cause, it was persistent over a sustained period of time. The other significant observation is that with only one exception, paleoliquefaction structures do not appear anywhere else higher in the section, despite
the ubiquity of meter-scale interbeds of fine- to medium-grained sandstone overlying fine-grained sandstone or siltstone throughout the section. Thus, the cause (or causes) of soft-sediment deformation appears to be temporally restricted to, at most, sand-rich interval I and enveloping siltstone units A and C, and it presumably ended by the time of deposition of sand-rich interval II (unit D). If it is assumed that the cause is earthquakes, then sand-rich intervals I and II record a transition from frequent seismic shaking to apparent seismic quiescence. Such an interpretation is consistent with previous suggestions that the end of mechanical stretching may have occurred near the base of the Johnnie Formation (Summa, 1993; Fedo and Cooper, 2001; Schoenborn et al., 2012). A ready alternative to a seismic trigger, however, is the effect of pressure contrasts from storm waves, which have also been shown to induce liquefaction and soft sediment deformation, including ball-and-pillow structure (Alfaro et al., 2002).

**Chemor stratigraphy**

A composite plot of $\delta^{13}C$ values of carbonate from the Johnnie Formation in southwest Laurentia (Verdel et al., 2011; this study) yields an overall pattern that is similar to profiles in Oman that contain the Shuram excursion, including a period of positive values as high as 4‰–6‰, rapid descent to values as low as –11‰ to –12‰, and a more gradual rise back to positive values (Fig. 15). The uniformly positive $\delta^{13}C$ values below the excursion in southwest Laurentia, generally of 1‰–3‰, invite detailed comparison with chemor stratigraphic profiles in the carbonate-rich Khufai Formation in Oman, which lies immediately below the type Shuram excursion. The stratigraphic thickness of units between the zero crossings of the Shuram excursion in Oman and southwest Laurentia are similar, ~500–700 m (Verdel et al., 2011). We therefore compared our profile to those from Oman without any modification to the vertical scaling (stratigraphic height), fixing the zero crossings at the base of the Shuram excursion at the same height. The Khufai sections in general are positive in $\delta^{13}C$ and show considerable variation, depending on the degree of diagenetic alteration. In least-altered sections (Mukhaibah Dome area), maximum values range up to 6‰, averaging 4‰–5‰ (Fig. 16A), i.e., considerably more positive than the Johnnie Formation profile. In more-altered sections (Buah Dome area; Fig. 16B), the profiles are quite similar to that of the Johnnie Formation. Given the close correspondence between the Johnnie profile and most of the Oman profiles (Fig. S3 [footnote 1]), we conclude that the data are consistent with, but do not absolutely demonstrate, temporal correlation between the upper part of the sub–Rainstorm Member Johnnie Formation (units H through L in the Spring Mountains) and the Khufai Formation.

The least-altered Khufai sections are generally considered to be representative of seawater carbon isotopic composition, defining a prolonged interval of $\delta^{13}C$ values in seawater near 6‰. Therefore, it seems clear that subsequent diagenesis was primarily responsible for reducing $\delta^{13}C$ values, in both Oman and the sub–Rainstorm Member Mount Schader section, by as much as 4‰–5‰. In the Pleistocene environment, such reduction has been shown to result from carbon isotopic exchange between carbonate beds and meteoric water, often resulting in $\delta^{13}C$ values decreasing stratigraphically upward at the scale of a few meters in beds exposed to erosion (Allan and Matthews, 1977, 1982; Quinn, 1991; Melim et al., 1995, 2001). The strong intrabed variations in $\delta^{13}C$ values in 1–2-m-thick carbonate intervals in the Johnnie Formation (Fig. 10) could potentially be explained by this mechanism, although $\delta^{13}C$ values of meteoric water at that time are poorly constrained and may not have been as
strongly negative as modern values. Further, the intrabed trends in δ¹³C values both increase and decrease downward, and there is no evidence of subaerial exposure on the tops of any of the beds. As with most Neoproterozoic carbonates, determining the mechanisms of depletion of δ¹³C values and their relationship to diagenetic textures and the biosphere is a difficult and controversial issue (Knauth and Kennedy, 2009; Derry, 2010a, 2010b; Grotzinger et al., 2011), and it is beyond the scope of this paper to resolve. One thing we can say, however, about the Mount Schader data set is that it displays no clear correlations between δ¹⁸O and δ¹³C (Fig. 11), as predicted by various isotopic exchange models (fig. 4 in Osburn et al., 2015). Despite this controversy, the good match between the type Johnnie Formation sub–Rainstorm Member section and the Khufai Formation supports the hypothesis that regardless of the origin of the anomalies, they nonetheless appear to be a useful correlation tool (Grotzinger et al., 2011). Tectonic reconstructions of the Neoproterozoic continent Rodinia put both the Shuram and the Johnnie formations roughly at the equator in Ediacaran time, but the two formations were located anywhere from 10,000 to 15,000 km away from each other (Li et al., 2008, 2013), making the isotopic correlation of the Shuram and sub-Shuram intervals all the more impressive.

One of the hallmarks of Neoproterozoic glacial cap carbonates is their frequent occurrence as thin, isolated intervals amid large thicknesses of enveloping strata that are entirely siliciclastic. Below unit H, there are two such isolated carbonate intervals, one in unit C and the other composing the entirety of unit F. Given their stratigraphic position between the Marinoan cap carbonate sequence and the base of the Cambrian, it is possible that either one of these units represents postglacial carbonate “rainout,” for example, as might be expected in the more southerly latitudes in the wake of the Gaskiers glaciation at 579 Ma (e.g., Pu et al., 2016). The generally positive δ¹³C values in the unit C and unit F carbonates, averaging between 1% and 2‰, argue strongly against either of these intervals representing a Gaskiers cap carbonate, which in Newfoundland yielded δ¹³C values of ~–8‰ to ~–2‰ (Myrow and Kaufman, 1999). Further, textural features widely described from cap carbonates (e.g., sheet cracks, tubes, teepee structures, etc.) are not observed in either of these intervals.

Subsidence Analysis

The substantial thickness of the Johnnie Formation, lack of evidence for unconformities in the Nevada sections, and the strengthened isotopic tie to the type Shuram excursion, motivate the hypothesis that the Noonday through lower Wood Canyon interval records continuous deposition through most or all of Ediacaran time. In the last section, backstripping and decomaption defined tectonic subsidence Y as a function of stratigraphic position S, independent of time. In this section, we model the element of time as exponential subsidence, assuming that Johnnie Formation and subsequent deposition of the passive-margin wedge occurred as a result of conductive cooling of rifted lithosphere. Subsidence analysis with well-defined ages at the Cambrian-Precambrian boundary (541 Ma) and at the base of Cambrian Age 5 (509 Ma) creates a considerably improved basis over previous studies for estimating stratigraphic age in Ediacaran strata by extrapolating the subsidence history back in time.

Regardless of the absolute elevation following mechanical extension of the lithosphere, once thermal subsidence begins, the elevation e of the surface, above its equilibrium value at t = 0, is closely approximated by:

\[ e(t) = E_r e_0 e^{-\frac{t}{t_v}}, \tag{1} \]

where \( E_r \) is the elevation of stretched lithosphere above its equilibrium depth at infinite time (or in the case of infinite stretching, the height of the ocean floor above the abyssal plains), t is time, \( t_v \) is the characteristic time (time at which \( e(t) = \frac{1}{e} e_0 \)) and
where $\beta$ is the stretching factor (Fig. 17; see Eqs. 10 and 11 in McKenzie, 1978). $E_0$ and $r$ are not parameters of interest when using subsidence as a chronometer, because we are attempting to use the late history of postrift subsidence, which is well dated, to constrain the earlier history of postrift subsidence, which is not. The simple exponential formula for elevation versus time $e(t)$ of Equation 1 is converted to subsidence depth $Y$ versus time by substituting $(E_0r - Y)$ for $e$, yielding:

$$Y(t) = E_0r(1 - e^{-t/\tau}).$$

In the case of mid-ocean ridges, where $\beta = \infty$ and $r = 1$, $E_0r$ is empirically shown to be within a few percent of 3.2 km (Parsons and Sclater, 1977). We note that this value does not correspond to the actual ridge elevation above the abyssal plain, which is much higher for oceanic crust less than 20 m.y. old. The characteristic time $\tau$, which depends on the thermal diffusivity and thickness of equilibrium lithosphere, shows somewhat greater variation depending on the ridge (±10% for the best-constrained ridges; table 1 in Parsons and Sclater, 1977).
Sclater, 1977), but a generally accepted range of values in subsidence analyses of passive margins is 50–65 m.y. (McKenzie, 1978; Allen and Allen, 2005). This corresponds to a “half-life” of thermal subsidence of 35–45 m.y. Even though this key parameter may vary significantly, we can estimate $\tau$ directly from our subsidence model, as an independent test of the hypothesis that the margin is in a state of exponential thermal subsidence, comparable to well-studied Mesozoic and Cenozoic examples. If our estimate of $\tau$ lies significantly outside the range of 50–65 m.y., it would falsify the thermal subsidence hypothesis.

Even though we do not know $E_Y$, estimation of $\tau$ and extrapolation of the curve back in time requires as few as two known elevation-time pairs, $(\epsilon_i, t_i$) and $(\epsilon_j, t_j$) (Fig. 17). Substituting these pairs into Equation 1, differenting the equations, and solving for $\tau$ yields:

$$\tau = \frac{(t_j - t_i)}{\ln \left( \frac{\epsilon_1}{\epsilon_2} \right)}$$

The differencing of the two equations eliminates $E_Y$, and hence the most important parameters in estimating both $\tau$ and the thermal subsidence curve itself are the elevation of two points relatively well separated in time from each other, and an estimate of zero elevation, i.e., where $\epsilon(\infty) = 0$ or the slope of $Y(t)$ is negligible. The thermal subsidence curve is then presumably applicable back in time to whatever point in the section at which we are still confident that the margin is in a state of pure thermal subsidence. As noted above, this level is probably no higher in the section than the lower part of the Johnnie Formation, and it may be much deeper, perhaps within the upper part of the underlying Pahrump Group.

Temporal constraints on the younger part of the subsidence curve are fairly similar to those used by Levy and Christie-Blick (1991), with the exception of their two oldest points, the base of the Cambrian Age 5 (approximately Middle Cambrian, 509 Ma; Walker et al., 2013) and the base of the Paibian (approximately Upper Cambrian, 497 Ma; Walker et al., 2013), which at the time were estimated to be 540 Ma and 523 Ma, respectively. Critically for this study, both the position and age of the Ediacaran-Cambrian boundary are well defined, lying within the Lower Member of the Wood Canyon Formation with an age of 541 Ma (Corsetti and Hagadorn, 2000). The 541 and 509 Ma constraints thus function as points $(\epsilon_1, t_1)$ and $(\epsilon_2, t_2)$, respectively, in our initial analysis, defining an exponential subsidence curve. As the oldest reliable temporally constrained points on the curve, they are the strongest constraints on extrapolating the curve back in time.

Points younger than 509 Ma are also well dated. These points clearly post-date the Sauk marine transgression, which marks a transition from predominantly siliciclastic to carbonate sedimentation, due to flooding of the craton through middle and late Cambrian time. Associated with the transgression, the average deposition rate $S(t)$ increases markedly from $\sim$20 m/m.y. from 541 to 509 Ma to $\sim$80 m/m.y. from 509 to 497 Ma (Table 4). Clearly, a fourfold increase in accumulation rate appears incompatible with any form of exponential subsidence. As explained below, the remarkable increase in subsidence rate owes its origin to the combination of sea-level rise and carbonate sedimentation, not renewed tectonism. The important point here is to note that the 541 and 509 Ma data points occur within the Lower Wood Canyon and Lower Carrara formations, respectively, both of which are shallow-water, mixed carbonate-siliciclastic facies associations that were probably deposited at similar points in global sea level. Both were deposited during highstand intervals relative to their transgressive substrates (the Stirling E Member and Zabriskie Formation, respectively). In the case of the Lower Member of the Wood Canyon Formation, the system evolved into a glacial drawdown of sea level (Smith et al., 2016). In the case of the lower Carrara Formation, sea level kept rising to a level that generally exceeded that of Ediacaran–early Cambrian time (Palmer, 1981).

The late subsidence history is characterized by very slow accumulation in Silurian and Early Devonian time ($<3$ m/m.y.; Table 4), and hence the difference between Silurian and Devonian values of $Y$ compared to those at 541 and 509 Ma provides firm estimates of $\epsilon_1$ and $\epsilon_2$. We note that with these constraints, the precise values of time and elevation for Paibian through Upper Ordovician strata provide little additional constraint on the form of the exponential subsidence curve.

**Temporal Model**

We present a temporal model of both observed subsidence $S(t)$ (i.e., stratigraphic thickness) and tectonic subsidence $Y(t)$ using a novel mode of presentation that orthogonally projects $Y(t)$ and $t(S)$ onto a graph of the numerically determined function $Y(S)$ (Fig. 18). In this approach, $Y(S)$ is plotted in the upper-left corner, $Y(t)$ is plotted in the upper-right corner, $r(S)$ is plotted in the lower-left corner, and $S(t)$ is plotted in the lower-right corner. The plot shows a simultaneous projection of $Y$ and $S$ onto their respective temporal models, graphically showing the influence of the slope of $Y(S)$ on the observed subsidence rates. The graph shows that, between 509 and 485 Ma, the increase in compressibility of the carbonate sediment [lower slope on $Y(S)$] combined with the accelerated schedule of subsidence caused by the flooding of the craton [higher slope on $Y(t)$] resulted in a dramatic increase in sediment accumulation rate [lower slope on $S(S)$ and higher slope on $S(r)$], even though exponential thermal subsidence was slowly decreasing. This result is critical, because it obviates the primary reason that most previous workers have cited in favor of Cambrian rifting along western Laurentia (e.g., Bond and Kominz, 1984; Levy and Christie-Blick, 1991; Yonkee et al., 2014).

**Parameter Estimates and Sensitivities**

Estimates of the exponential time constant $\tau$ vary according to two main uncertainties: (1) the sediment grain density assumed in our delithification model, and (2) whether or not a thick substrate of Pahrump Group strata is present at depth beneath the exposed Spring Mountains section. We calculated values
of \( \tau \) for values of tectonic subsidence \( Y \) in a series of models that encompass these parameter variations (Table 5). In addition, we defined \( Y \) according to two different assumptions for the point at which mechanical stretching ends and purely thermal subsidence begins, where \( Y = 0 \) (i.e., \( e = E/e_i \)). One is at the lowest exposed stratum (base of unit A), and the other is within unit C, above the youngest ball-and-pillow structure at the top of unit B, assuming seismic shaking ended near this point. In Table 5, models with no Pahrump Group substrate are designated \( Y_{\text{ns}} \), and those that include the substrate are designated \( Y_{\text{ws}} \); intermediate-density models contain no additional subscript, and low- and high-density models are also subscripted “low” and “high,” respectively. Models with \( ** \) define \( Y = 0 \) within unit C, and models with no superscript assume \( Y = 0 \) at the base of unit A. We defined the value of \( Y \) for which \( e = 0 \) to be the average of \( Y(393 \text{ Ma}) \) and \( Y(383 \text{ Ma}) \), designated \( Y(c.388) \) in Table 5. The results are insensitive to this choice because there is so little variation in \( Y \) between 444 and 383 Ma. We cannot choose the next younger point in the subsidence profile (359 Ma), because it clearly reflects the onset of subsidence associated with Antler foredeep sedimentation.

The contrast in \( \tau \) between models \( Y_{\text{ns}} \) and \( Y_{\text{ws}} \) is only 3 m.y., with \( \tau = 55 \) and 52 m.y., respectively. As expected from Equation 4, the definition point of \( Y = 0 \) has no effect, because we define \( e_1 \) and \( e_2 \) on the basis of differences in \( Y \) values late in the subsidence history. For models with no substrate, varying the density between \( Y_{\text{ns, low}} \) and \( Y_{\text{ns, high}} \) (corresponding to the assumption of high and low sediment grain density, respectively) has a substantial effect on \( \tau \), which ranges from 42 to 65 m.y., respectively. For models \( Y_{\text{ws, low}} \) to \( Y_{\text{ws, high}} \), the sensitivity is even greater, with \( \tau \) varying from 36 to 63 m.y., respectively. Clearly, the upper part of these ranges accords with subsidence patterns in Mesozoic and Cenozoic basins. Despite the nearly 30 m.y. variation in \( \tau \) among these models, we note that there is relatively little variation in the modeled age and duration of the Shuram excursion (Table 5). Among this suite of models, the onset varies by 12 m.y. (from 569 to 581 Ma), the
termination varies by 9 m.y. (from 566 to 575 Ma), and the duration varies by 3 m.y. (from 3 to 6 m.y.).

A further consideration in estimating $\tau$ is the fact that because the Sauk transgression was well under way by 509 Ma, relative sea level may have been slightly higher than at 541 Ma. To the extent that it was, a significant systematic transgression was well under way by 509 Ma, relative sea level may have been 3 m.y. (from 3 to 6 m.y.).

For example, for the model with no count for a change in sea level (+50 m in $Y$ subsequent cratonic flooding events. The resulting subsidence model for the point at 541 Ma, minimizing both the amount of extrapolation back in time, and the range of values indicated by models of Mesozoic and Cenozoic basins both suggest a value toward the upper end of this range. We further note that the earliest empirical fits to long-term seafloor subsidence data suggested a value of 62.8 m.y. (Eq. 22 in Parsons and Sclater, 1977; Table 6 in McKenzie, 1978).

| Model   | $Y(541)$ | $Y(509)$ | $Y(393)$ | $Y(383)$ | $Y(c. 388)^b$ | $\tau$ (m.y.) | SE begin$^d$ (Ma) | SE end$^d$ (Ma) |
|---------|----------|----------|----------|----------|---------------|---------------|-----------------|-----------------|
| $Y_m$   | 2009     | 2327     | 2713     | 2743     | 2728          | 55            | 578             | 573             |
| $Y_m$$^d$ | 1646     | 1964     | 2350     | 2380     | 2365          | 55            | 578             | 573             |
| $Y_{ws}$ | 2757     | 3029     | 3334     | 3358     | 3346          | 52            | 575             | 570             |
| $Y_{ws, low}$ | 1844   | 2125     | 2367     | 2380     | 2374          | 42            | 574             | 569             |
| $Y_{ws, high}$ | 2009 | 2327     | 2713     | 2743     | 2728          | 55            | 578             | 573             |
| $Y_{ws, low}$$^d$ | 2173   | 2528     | 3058     | 3105     | 3082          | 65            | 581             | 575             |
| $Y_{ws, high}$$^d$ | 1500 | 1781     | 2023     | 2036     | 2030          | 42            | 574             | 569             |
| $Y_{ws, low}$$^d$ | 1646 | 1964     | 2350     | 2380     | 2365          | 55            | 578             | 573             |
| $Y_{ws, high}$$^d$ | 1791 | 2146     | 2676     | 2723     | 2700          | 65            | 581             | 575             |
| $Y_{ws, low}$ | 2475 | 2710     | 2871     | 2878     | 2875          | 36            | 569             | 566             |
| $Y_{ws}$ | 2757     | 3029     | 3334     | 3358     | 3346          | 52            | 575             | 570             |
| $Y_{ws, high}$ | 3037 | 3347     | 3795     | 3836     | 3816          | 63            | 578             | 573             |

$^d$Units for $\tau$ are millions of years (m.y.). Units for all $Y$ values are in meters (m).

$^b$Mean value of $Y(393/\text{Ma})$ and $Y(383/\text{Ma})$.

$^d$SE—Shuram excursion.

$^d$**$Y$ values were adjusted by assuming a zero datum that represents a specific point in the stratigraphic column inferred to represent cessation of mechanical stretching and inception of passive-marginal thermal subsidence.

Termination of the Shuram excursion in Oman (Le Guerroué et al., 2006b), which has been called
into question on the basis that the Khufai/Shuram interval was probably not deposited on a thermally subsiding continental shelf (Bowring et al., 2007).

In sum, because the timing of the Shuram excursion is within ~0.5 \( \tau \) of 541 Ma, varying parameters in the exponential subsidence model yields variations in our estimate of age and duration of the Shuram excursion of just a few million years. The fact that a fairly broad range of parameters leads to estimates of the end of the Shuram excursion centered on 579 Ma suggests that the valleys incised into the Rainstorm Member are indeed a manifestation of the Gaskiers glaciation at equatorial latitudes. To conclude otherwise strains credibility, because Johnnie/Stirling sequence architecture is relatively uneventful for 400–500 m both above and below the Rainstorm Member (Stirling Member A/B and Johnnie units H through L, respectively). If incision was unrelated to the Gaskiers glaciation, this requires: (1) that the most dramatic stratigraphic event in the Johnnie/Stirling interval was close in time, but unrelated to, glaciation; and (2) that the Gaskiers glaciation itself had virtually no impact on the section. In essence, the subsidence analysis provides a relatively coarse estimate of age that “registers” the section with possible correlatives elsewhere. The detailed stratigraphy then fine tunes the age estimate based on a specific correlation with well-dated events elsewhere, in this case, shelf-incision and the Gaskiers glaciation.

The overall consistency of exponential subsidence models with the hypothesis that incision of the Rainstorm Member shelf is an expression of the Gaskiers glaciation is supported by the following evidence. The table below shows the model ages for various units, with the top of the unit as the age of interest. The table includes the ages for different values of \( \tau \).
the Gaskiers glaciation suggests that modeled ages of other horizons in the Johnnie/Stirling interval may also be accurate to within a few million years. The overall accuracy of this model can be further tested by assessing how well it estimates the age of the lowermost Johnnie and Noonday interval. As noted above in our discussion of the possible correlation of unit A with the Noonday Formation, we would expect the age of this unit to be close to the age of the base of the Noonday Formation, or 635 Ma (Petterson et al., 2011). The ranges of modeled ages for the base of unit A are 639–608 Ma (Table 6), with the “midrange” model shown in Figure 18B predicting an age of 628 Ma. As shown in Figure 18D, linear extrapolation below the deepest exposed strata of unit A, assuming a linear deposition rate, would require only an additional 288 m of “subunit A” strata to bring the section to the base of the Noonday Formation and the Ediacaran Period. This thickness plus the 125 m thickness of unit A yields a total thickness of 413 m, which is consistent with maximum known thicknesses of the Noonday Formation (Petterson et al., 2011). The apparent success of exponential subsidence models in predicting the age of both the Gaskiers event and the base of the Ediacaran Period at their most likely stratigraphic levels supports the hypothesis that Ediacaran deposition on the southwest Laurentian margin was largely continuous, and that the Noonday through Wood Canyon interval in its thickest, most basinal exposures does not contain unconformities with significant depositional hiatuses.

CONCLUSIONS

Lithostratigraphic and chemostratigraphic details of the Johnnie Formation at its type locality in the northwest Spring Mountains of southern Nevada provide a basis for regional lithostratigraphic correlation, global chemostratigraphic correlation, and subsidence analysis of the southwest Laurentian continental margin. The regional lithostratigraphy of Ediacaran through Cambrian Age 4 strata defines seven sand-rich intervals separated by siltstone- and carbonate-rich intervals, the upper two of which are the Sauk I and Sauk II sequences of Cambrian age (Palmer, 1981). The great overall thickness of the Johnnie Formation at its type locality (~1800 m), and the apparent absence of subaerial exposure surfaces or other evidence of erosion that are well expressed in more cratonic sections, such as the Nopah Range section, support the hypothesis that Ediacaran deposition was nearly continuous, and that the Noonday through Wood Canyon interval in its thickest, most basinal exposures does not contain unconformities with significant depositional hiatuses.

ACKNOWLEDGMENTS

We are grateful to Gillian Anderson, Leah Sabbath, Fanfeng Wu, and the late Lindsey Hedges for assistance in the field and laboratory, and to Associate Editor Christopher J. Spencer and reviewer Tony Prave for insightful and constructive reviews. This material is based upon work supported by the National Science Foundation Graduate Research Fellowship Program under grant 1144469 awarded to R. Witkosky, and grant EAR 1451055 awarded to B. Wernicke.

APPENDIX. DESCRIPTION OF MAP UNITS

Descriptions apply to geologic maps and stratigraphic columns shown in Figures 3, 4, 6, and 7.

| Unit Code | Description |
|-----------|-------------|
| Qt:       | Interbedded limestone, siltstone, sandstone, and shale. |
| Qa:       | Alluvium and colluvium in active/ephemeral channels and piedmont-forming slopes. |
| QZa:      | Stirling Formation, A Member (labeled “Za” on maps). Very pale-orange, grayish-black-weathering, medium-grained orthoquartzite, laminated to massive, medium to thick bedded, with trough cross-stratification. Contains some interbedded carbonate-cemented sandstone. In places, bedding is destroyed by secondary brecciation and reparation, forming irregular dark-weathering masses. Unit forms resistant ridges relative to underlying Johnnie Formation. |

Carbon isotopic data from sub–Rainstorm Member (sub–Shuram excursion) units in the Mount Schader section are generally positive and support correlation of Johnnie Formation units H through L with the Khufai Formation in Oman, but they do not require it. If correlative, the Mount Schader section would provide the first confirmation of an extended period (represented by 300–400 m of section) of positive δ 13C values prior to the Shuram excursion in both Oman and Nevada.

The Gaskiers glaciation marks the beginning of widespread preservation of macroscopic Ediacaran animals (Xiao et al., 2016), and the Shuram excursion is the largest known carbon isotopic excursion in the geological record. A central issue in animal evolution is thus whether or not the Shuram excursion was approximately synchronous with the Gaskiers event, because it suggests that the Shuram excursion, whatever its cause, was genetically related to creating a surface environment that could support the metabolic requirements of macroscopic animals. A second consequence of Shuram-Gaskiers correlation is that it places the transition from diverse, ornamented acritarchs to a lower-diversity, unornamented assemblage in synchrony with the appearance of macroscopic animals, rather than at some later time. The issue is addressable in southwest Laurentia, to the extent that deposition of Johnnie Formation and related strata occurred more-or-less continuously on a thermally subsiding passive margin.

Based on this assumption, subsidence analysis strongly suggests that the end of Johnnie Formation deposition, at the time of valley incision and subsequent fill with the conglomeratic member, was correlative with the Gaskiers glaciation at 579 Ma. The analysis also suggests that the onset of the Shuram excursion near the base of the Rainstorm Member occurred at ca. 585 Ma. The implied 6 m.y. duration of the Shuram excursion is consistent with paleomagnetic and other proxies from sections in Laurentia and Australia. The subsidence analysis further indicates that if the assignment of the Gaskiers event to uppermost Johnnie time is correct, then the base of the Johnnie Formation is ca. 630 Ma. If so, then the Johnnie through lower Wood Canyon interval in the Spring Mountains represents a relatively complete, 3000-m-thick section that records most or all of Ediacaran time.
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