Geochronologic and stratigraphic constraints on the Mesoproterozoic and Neoproterozoic Pahrump Group, Death Valley, California: A record of the assembly, stability, and breakup of Rodinia

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ABSTRACT

The Pahrump Group in the Death Valley region of eastern California records a rich history of Mesoproterozoic to Neoproterozoic tectonic, climatic, and biotic events. These include the formation, stability, and onset of rifting of the Rodinia supercontinent, two potentially low-latitude glaciations correlative with global "snowball Earth" glacial intervals, and the onset of complex microbiota (e.g., testate amoebae). Poor direct age control, however, has significantly hindered the progress of understanding of these important stratigraphic units. New LA-ICPMS (laser ablation–inductively coupled plasma mass spectrometry) detrital zircon data from clastic units directly overlying a major unconformity within the Mesoproterozoic Crystal Spring Formation provide a maximum depositional age of 787 ± 11 Ma for the upper member of the Crystal Spring Formation. This unconformity, representing a duration of ≥300 Ma, is now recognized in sedimentary successions across southwestern Laurentia. These new age data, in addition to the distinct stratigraphic style above and below the unconformity, result in the proposed formal stratigraphic revision to elevate the upper member of the Crystal Spring Formation to the Neoproterozoic Horse Thief Springs Formation and separate it from the remainder of the underlying Mesoproterozoic Crystal Spring Formation (ca. 1100 Ma). New age relations and revised stratigraphic nomenclature significantly clarify stratigraphic and tectonic correlations and imply ca. 1250–1070 Ma assembly, 1070–780 Ma stability, and 780–600 Ma breakup of the supercontinent Rodinia along the southwestern Laurentian margin.

INTRODUCTION

The Mesoproterozoic–Neoproterozoic Pahrump Group of the Death Valley area, California (Figs. 1 and 2), is an important geologic record for understanding Rodinia reconstructions, extreme global climate change, and biotic evolution, hence its geology is well known internationally. These well preserved, easily accessible, and regionally exposed strata are of particular interest because they record: extension attributed to both the culmination and rifting of the supercontinent Rodinia (Wright et al., 1974; Labotka et al., 1980; Miller, 1987; Timmons et al., 2001, 2005; Petterson, 2009; Petterson et al., 2011a); the evolution of the western Laurentian margin (Prave, 1999; Fedo and Cooper, 2001); two low-latitude glaciations and their aftermath (Troxel, 1982; Miller, 1985; Prave, 1999; Abolins et al., 2000; Corsetti and Kaufman, 2003); and windows into prokaryotic and eukaryotic evolution prior to the Cambrian explosion (Cloud et al., 1969; Licari, 1978; Horodyski and Mankiewicz, 1990).

Despite extensive research on Pahrump Group stratigraphy (e.g., Hewett, 1940, 1956; Wright, 1949; Wright and Troxel, 1956, 1966; Troxel, 1967; Gutstadt, 1968, 1969; Cloud et al., 1969; Maud, 1979, 1983; Roberts, 1982; Wright and Prave, 1993; Prave, 1999; Petterson et al., 2011a, 2011b), fundamental disagreements about formal stratigraphic boundaries, inferred age constraints, and correlations across the outcrop area of the Pahrump Group remain unresolved (e.g., Mbuyi and Prave, 1993; Prave, 1999; Corsetti et al., 2007; Mrofka and Kennedy, 2011; Petterson et al., 2011a, 2011b). These disagreements significantly hinder regional correlations and paleogeographic reconstructions involving this key region.

This paper presents U-Pb maximum depositional ages from detrital zircon grains, which constrain the magnitude of a major unconformity within the Pahrump Group. We also present stratigraphic revisions, which significantly clarify the correlations both within the Pahrump Group strata and also to stratigraphic successions across, and beyond, the southwestern Laurentian margin. Thus, these new geochronologic and stratigraphic results advance the essential time-stratigraphic framework for understanding the Laurentian and global Mesoproterozoic and Neoproterozoic Earth system.

GEOLOGIC BACKGROUND

Previous Geochronologic Constraints

Chronologic constraints in the Pahrump Group are sparse, and most ages have been determined using indirect lithostratigraphic and chemostratigraphic correlations to similar units and geologic events, both regionally and globally (e.g., Prave, 1999; Dehler et al., 2001, 2011; Timmons et al., 2005; Corsetti et al., 2007; Dehler, 2008). However, without direct geochronologic constraints, these correlations remain tenuous. Currently the only direct radiometric constraints on the Pahrump Group are two U-Pb baddeleyite ages of 1069 ± 3 Ma and 1087 ± 3 Ma from diabase sills in the lower and middle Crystal Spring Formation (Heaman and Grotzinger, 1992; see Fig. 2). Wright (1968) and Hammond (1983, 1986) suggested intrusion of these sills to have occurred while sediments of the Crystal Spring Formation were wet and unconsolidated, thus

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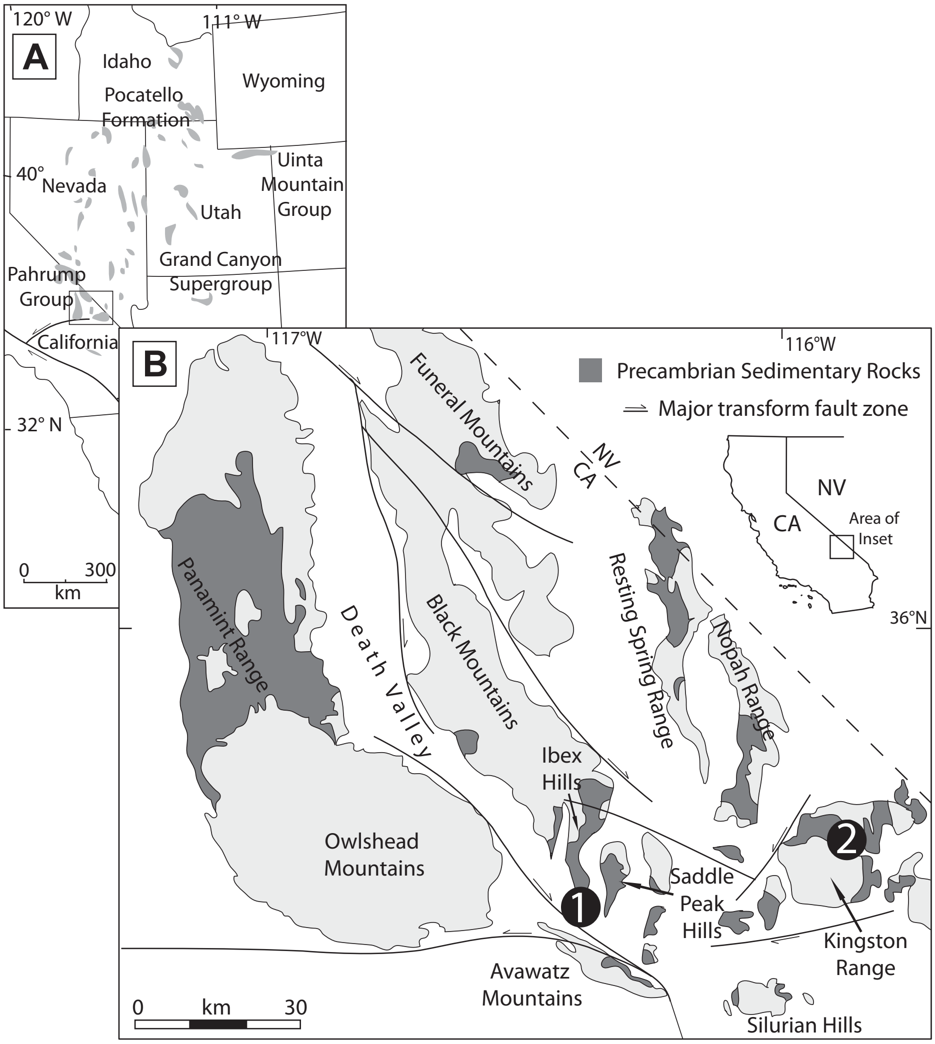

Figure 1. (A) Map of the southwestern United States showing location and extent of Neoproterozoic sedimentary successions (modified after Stewart et al., 2001; Lund, 2008). (B) Distribution of Proterozoic sedimentary rocks in the southern Death Valley region (CA—California; NV—Nevada). Geology modified from Workman et al. (2002) and Petterson (2009). Localities sampled and discussed in text are indicated by numbers: 1—southern Ibex Hills/Saratoga Spring (reference locality of the Horse Thief Springs Formation; sample locality for samples K03DV10, K03DV11, 12RMSS5); 2—Beck Canyon in the Kingston Range (type locality of the Horse Thief Springs Formation; sample locality for sample 4CD11).

these ages approximate a depositional age. Only two other age constraints from the Death Valley Proterozoic section are reported: the presence of the trace fossil *Treptichnus pedum* approximately marking the Neoproterozoic-Cambrian boundary in the Wood Canyon Formation (Corsetti and Hagadorn, 2000), ~3 km upsection from the top of the Pahrump Group, and a single detrital zircon grain from the Johnnie Formation which yielded a U-Pb CA-TIMS (chemical abrasion–thermal ionization mass spectrometry) age of 640.33 ± 0.09 Ma (Verdel et al., 2011). This single grain, however, does not represent a reproducible population from which age constraints can be reliably determined (see Dickinson and Gehrels, 2009) and would represent the maximum depositional age, at best.

Historical Usage of Stratigraphic Nomenclature

Hewett (1940) originally defined the "Pahrump series" as consisting of three formations: a lower siliciclastic-carbonate unit, the "Crystal Spring formation"; a middle carbonate unit, the "Beck Spring dolomite"; and an upper siliciclastic unit, the "Kingston Peak formation." The initial definition of stratigraphy by Hewett (1940) was broad-brush in nature, and only gave reference to type localities, with no type sections described. The first detailed stratigraphic studies within the Pahrump series were conducted on the Crystal Spring Formation (Wright, 1949, 1952), and on representative sections of the three formations within the Pahrump series (Hewett, 1956). Applications of member-scale subdivisions for the Pahrump Group are common in historical literature and are generally presented informally with little deference given to precedence or consistency. Wright (1949, 1952) first proposed informal subdivisions for the Crystal Spring Formation. Modifications of these subdivisions were used by several subsequent workers, some of whom were students of Wright (e.g., Roberts, 1974a, 1974b, 1976, 1982; Maud, 1979, 1983). Historical nomenclature for the Pahrump Group, with emphasis on the Crystal Spring Formation member subdivisions and significant formation- or group-level revisions, is presented in Table 1.

Crystal Spring Formation

The Crystal Spring Formation, of particular interest in this study, was previously studied in stratigraphic detail, including the assignment of type sections and informal member divisions (Wright, 1949, 1952; see Table 1). Informal member names used for the lower and middle parts of the Crystal Spring Formation, in ascending stratigraphic order, include: the arkosic sandstone, feldspathic sandstone, mudstone, dolomite, algal dolomite-siltstone, and chert members (*sensu* Wright, 1968; Roberts, 1974a, 1974b, 1976, 1982), all intruded by diabase sills. The "upper units" (*sensu* Wright, 1968; Roberts, 1974a, 1974b, 1976, 1982), also referred to as the upper member by contemporary workers, consists of five 10–100-m-scale siliciclastic-carbonate cycles overlain by a transitional interval of mixed carbonate-siliciclastic meter-scale cycles (Maud, 1979, 1983). Stratigraphic dissimilarities between the lower and middle parts of the Crystal Spring Formation compared with the upper units were implied by early work (Table 1; Roberts, 1974a, 1974b, 1976, 1982; Maud, 1979, 1983), and an unconformity at the base of the upper units of the Crystal Spring Formation was suggested by both Maud (1979, 1983) and Mbuyi and Prave (1993). For the purpose of consistency with more recent publications (e.g., Heaman and Grotzinger, 1992; Prave, 1999), we generally refer to the arkosic sandstone, feldspathic sandstone, and mudstone units as the "lower member"; the dolomite, algal dolomite-siltstone, and chert members as the "middle member"; and the upper units as the "upper member" (see Table 1).

Lower and Middle Members of the Crystal Spring Formation

The lower member of the Crystal Spring Formation (≤660 m thick) unconformably overlies crystalline basement and comprises a basal quartzite-granite conglomerate (arkose member of Wright [1952, 1968]), which fines upward into sandstone and ultimately mudstone. These deposits are interpreted to represent an earlier south-flowing fluvial system that later shifted to an intertidal-deltaic system coming from a southern source (Roberts, 1974a, 1974b, 1976, 1982).

The middle member of the Crystal Spring Formation (≤397 m thick) comprises thick microbially laminated limestone and dolomitic limestone overlain by clastic strata, with the limestone units dominating in the north and the siliciclastic strata prominent in the south. These strata record shallow-water carbonate sedimentation prograding from the northern margin of the basin as clastic material continued to be fed into the basin from southerly sources (Roberts, 1974a, 1974b, 1976, 1982).

Upper Member of the Crystal Spring Formation

An unconformity marks the boundary between the middle and upper Crystal Spring Formation. This unconformity was originally recognized by Maud (1979, 1983) in the Saddle Peak Hills where the upper member rests in angular discordance on the middle member. Maud (1983) also noted that the conglomerate/breccia at the base of the upper member has clasts derived almost entirely from the "chert member" of the underlying middle Crystal Spring Formation, implying that the middle member had to have been uplifted and reworked to generate the overlying deposit. Mbuyi and Prave (1993) further described this unconformity on a regional scale and recognized that conglomerates above the unconformity occasionally contain diabase clasts (e.g southern Ibex Hills/Saratoga Spring section) indicating significant unroofing and missing time, although no inference was made as to the actual duration represented by this surface. The unconformity is well exposed in the southern Ibex Hills (near Saratoga Spring) and the northern Saddle Peak Hills, where the basal clastic unit is overlain by the first major carbonate unit of the upper member, making together the "A unit" of Maud (1979, Table 1, 1983).

Overlying the unconformity are six regionally traceable cyclic siliciclastic-carbonate intervals, informally described as "A through F units" by Maud (1979, 1983). The basal deposits (lower part of A unit) comprise several meter-scale cycles of clast-supported conglomerate and/or breccia that fine upward into coarse-sand to granule sandstone (Figs. 3A–3D). The clast composition is almost entirely chert and hornfelsic siltstone derived from the underlying "chert member" (*sensu* Roberts, 1974a, 1974b, 1982) of the middle member of the Crystal Spring Formation (Maud, 1979, 1983), except where the basal unit rests directly on the diabase sill and has dominantly clasts of diabase (A.R. Prave, 2003, personal commun.).

Each of the lower five cyclic intervals (A–E units, out of 6 total intervals) contains a lower

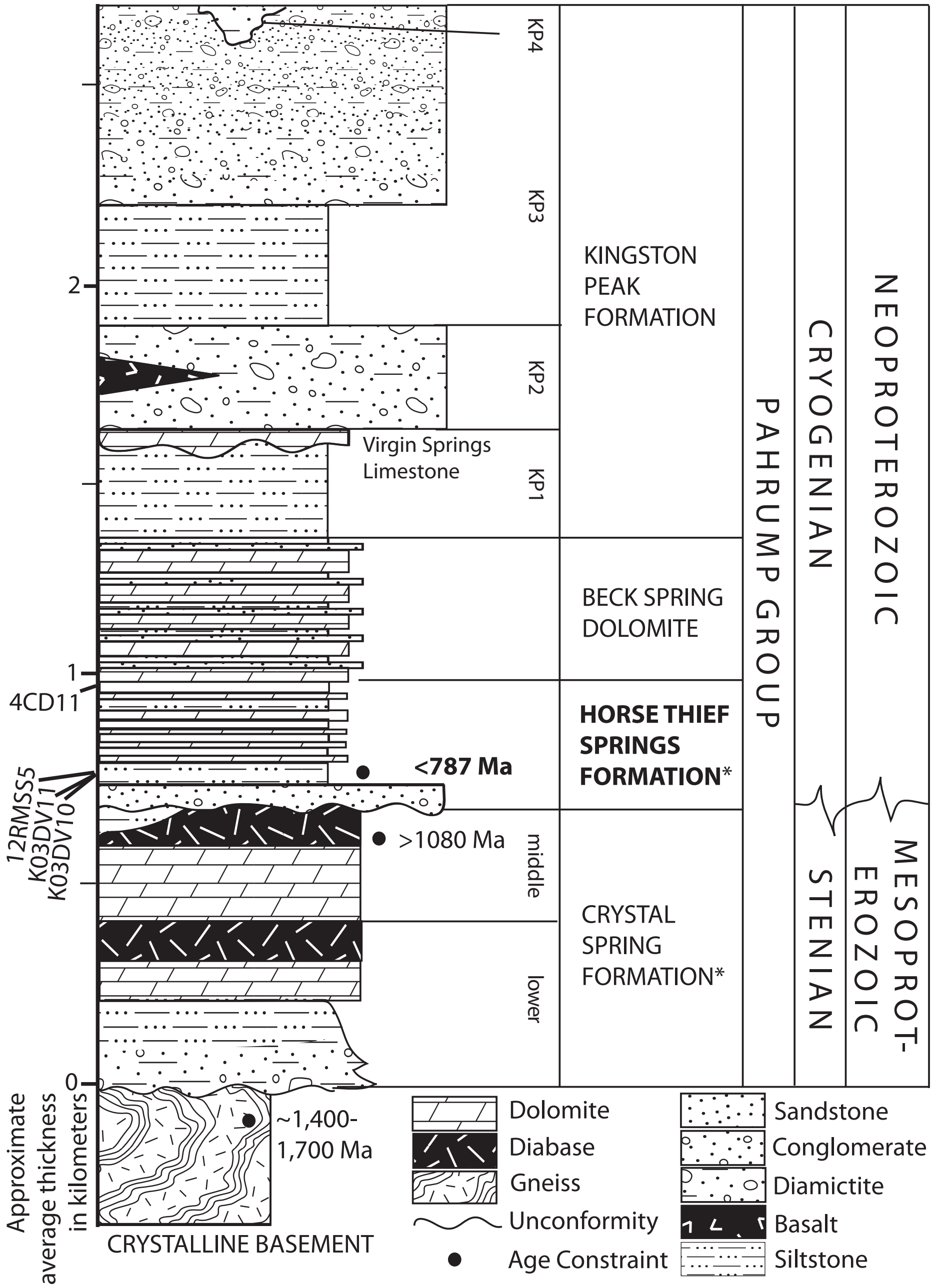

Figure 2. Generalized stratigraphy of the Pahrump Group in the Death Valley region. Sample numbers indicate the stratigraphic position of samples discussed in this study. Asterisks indicate stratigraphic nomenclature to be revised in this publication. New age constraint and stratigraphic revisions to be discussed are labeled in bold. Radiometric age constraints for the lower and middle Crystal Spring Formation from Heaman and Grotzinger (1992). Basement age constraints from Lanphere et al. (1964), Labotka et al. (1980), Barth et al. (2001, 2009), and Iriondo et al. (2004).

TABLE 1. HISTORICAL USAGE OF STRATIGRAPHIC NOMENCLATURE

Hewett, 1940; 1956	Wright, 1952	Kupfer, 1960	Jennings et al., 1962	Roberts, 1974a; 1974b	Maud, 1979; 1983	Grotzinger, 1992	This Study	
Pahrump series		Pahrump group	Pahrump Group					
Kingston Peak formation	conglomeratic quartzite mbr.	upper Kingston Peak formation	Kingston Peak Formation				Kingston Peak Formation	KP4
								KP3
								KP2
	green quartzite mbr.	lower Kingston Peak formation						KP1
Beck Spring dolomite		Beck Spring dolomite	Beck Spring Dolomite				Beck Spring Dolomite	
Crystal Spring formation	upper sedimentary units	Crystal Spring formation	Crystal Spring Formation	upper units	F E D C B A units	upper mbr.	Horse Thief Springs Fm.*	F E D C B A units
	massive chert mbr.			chert mbr.	chert mbr.	middle mbr.	Crystal Spring Formation	middle mbr.
	carbonate mbr.			algal mbr.	carbonate member			
				dolomite mbr.				
	fine grained quartzite mbr.			mudstone mbr.		lower mbr.		lower mbr.
	purple shale mbr.			felds. ss. mbr.				
	feldspathic quartzite mbr.			arkose mbr.				

*Nomenclature formally defined in this study
Abbreviations: mbr.—member; Fm.—Formation; felds.—feldspathic; ss.—sandstone.

siliciclastic component, ranging in grain size from silt to gravel, and an upper carbonate component, with each cycle exhibiting distinctive characteristics. A, B, and E units are on the order of tens of meters thick, depending on the location in the basin. C, D, and F units range in thickness from less than 10 m in the west-southwest to ~160 m in the east-northeast (this study; see Figs. DR1, DR2 in the GSA Data Repository[1]). The D unit and basal E unit are not present in the southern Saratoga Spring area and instead are represented by an intraformational unconformity that places the upper E unit on the C unit (Maud, 1979, 1983).

The sandstone units of the lower five cycles are typically quartzose and the conglomerate facies are typically dominated by chert clasts from the underlying middle member of the Crystal Spring Formation (this study; see Figs. 3C, 3D, 3F). An exception is the D unit, where there is a prominent arkosic wedge indicating input and likely uplift from a granitic source to the southwest. The siliciclastic units generally thicken to the east and south whereas grain size in these units coarsens to the south-southwest and becomes finer northward, indicating an upland to the south-southwest and maximum subsidence rate in the east (Maud, 1979, 1983).

The carbonate units overlying the clastic units in the lower five cyclic intervals generally thicken to the southwest, are individually unique in sedimentary character, and are commonly stromatolitic. These dolomite intervals include all or some of the following: cryptalgal laminations, allochemical grainstone, oolites, oncolites, cherty beds, and stromatolites including *Conophyton* and *Baicalia* (Maud, 1979, 1983; S.M. Awramik, 2012, personal commun.).

The capping F unit is an interval of locally developed, thinner, stacked meter-scale cycles consisting of siliciclastic units overlain by dolomitic units. Siliciclastic units are generally thinner and coarser grained in the southwestern sections and exhibit internal channel scour and fill, while eastern sections are thicker and finer grained (generally bedded to laminated, silt to fine sand). Dolomitic units are thin (0.1–2 m) and commonly contain sand- and granule-sized siliciclastic grains, rip-up clasts, and cryptalgal or stromatolitic lamination. The tops of the dolomitic units also commonly exhibit paleokarst and grikes (solution fissures) (Maud, 1979, 1983). The number of siliciclastic-carbonate cycles dramatically increases from the east (where only a few are present in the Kingston Range) to the west (where several tens of cycles are present). These meter-scale carbonate-siliciclastic cycles are locally only traceable over short distances (tens to hundreds of meters) and transition either gradationally (in western sections) or abruptly but conformably (in eastern sections) into the overlying Beck

[1]GSA Data Repository item 2014108, figure DR1–DR6 and table DR1–DR3, is available at http://www.geosociety.org/pubs/ft201X.htm or by request to editing@geosociety.org.

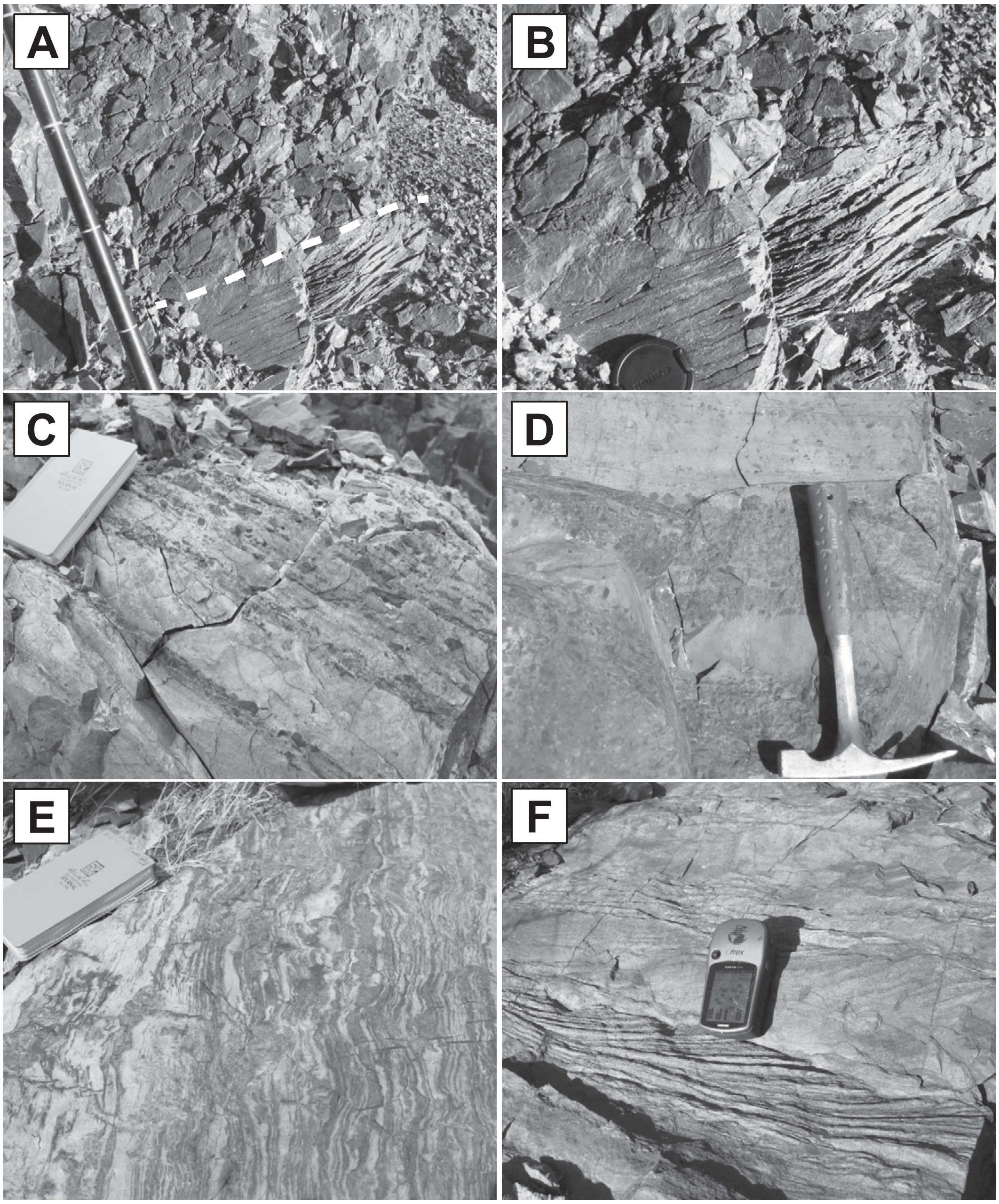

Figure 3. Outcrop photographs. (A) Unconformity at the base of the upper member of the Crystal Spring Formation (revised herein to Horse Thief Springs Formation) from the southern Ibex Hills. White line shows location of unconformity, with middle member of the Crystal Spring Formation hornfelsic siltstone to the lower right, upper member of the Crystal Spring conglomerate above (lines on staff are 10 cm apart). (B) Detail of unconformity surface in the southern Ibex Hills, showing metamorphosed siltstone below and clasts of siltstone above the unconformity (stratigraphic facing to the upper left). (C) Conglomeratic sandstone near the base of the upper member of the Crystal Spring Formation (Horse Thief Springs Formation) from the Saddle Peak Hills; clasts comprise metamorphosed siltstone and chert derived from the middle Crystal Spring Formation (stratigraphic facing to the upper right). (D) Siltstone and fine-grained sandstone with interbedded chert and cherty siltstone–pebble conglomerate near the base of the upper member of the Crystal Spring Formation (Horse Thief Springs Formation) in the southern Ibex Hills. (E) Enterolithic folding (soft-sediment deformation resulting from dissolution of evaporites) in siltstone and fine sandstone of the upper member of the Crystal Spring Formation, A unit, from Beck Canyon in the Kingston Range (field notebook for scale is 19 cm by 12 cm; stratigraphic facing to the right). (F) Trough cross-bedding in the upper member of the Crystal Spring Formation (Horse Thief Springs Formation), A unit, from Beck Canyon in the Kingston Range (GPS unit for scale is 11cm in length).

Spring Dolomite. The transitional nature of this contact is particularly well presented in sections near Saratoga Spring in the southern Ibex Hills, where the exact location of the base of the Beck Spring Dolomite is unclear.

Maud (1979, 1983) proposed four depositional models to explain the collective stratigraphic and sedimentologic data from this unit: (1) intertidal model; (2) barrier beach model; (3) tectonic model; and (4) open clastic shoreline model. The lattermost of these models was preferred because the majority of sandstone units were interpreted as beach sands and the carbonates were considered subtidal due to lack of exposure indicators as well as their lateral continuity.

METHODS

Detrital Zircon Analysis

Sampling

Samples from the upper member of the Crystal Spring Formation were collected for detrital zircon analysis to test the hypothesis that the upper member is indeed a different and significantly younger unit than the lower and middle members of the Crystal Spring Formation. Samples were collected from two separate localities: the first (samples K03DV10, K03DV11) near the parking area at the end of Saratoga Spring road and the second (sample 12RMSS5) from the west-trending ridge ~1 km north of the first sample locality (see Table 2, Fig. 1). A single detrital zircon age is reported from the top of the upper member of the Crystal Spring Formation (sample 4CD11) in Beck Canyon of the Kingston Range (see Fig. 1).

TABLE 2. DETRITAL ZIRCON SAMPLE LOCALITIES AND DESCRIPTIONS OF SAMPLES

Sample number	Formation	Member	Locality	GPS Northing*	GPS Easting*	Description
4CD11	Horse Thief Springs Formation	F unit	Crystal Spring, Kingston Range	0592342	3961793	Red sandstone
12RMSS5	Horse Thief Springs Formation	A unit, second conglomerate	Saratoga Spring, southern Ibex Hills	0552634	3949335	White quartzite
K03DV11	Horse Thief Springs Formation	A unit, third conglomerate	Saratoga Spring, southern Ibex Hills	0552487	3948640	White quartzite
K03DV10	Horse Thief Springs Formation	A unit, second conglomerate	Saratoga Spring, southern Ibex Hills	0552484	3948639	Hornfels clasts in white quartzite matrix

*GPS data given in Universal Transverse Mercator Zone 11S, North American Datum 27.

Preparation

Samples were prepared for analysis using standard crushing and mineral separation techniques at Boise State University (Idaho) and University of Arizona mineral separation facilities. Samples were washed and crushed using hammer and steel plate or mechanical chipper, then powdered in a disc mill and sieved to 60 mesh. Sieved sand grains were run across a water table for relative density separation and sample was collected at three intervals. The heaviest washed sample was then oven dried and subjected to initial Frantz magnetic separation, and heavy liquid separation using methylene-iodide (density ρ = 3.32 g/cm^3) heavy liquid. Heavy mineral separates were then subjected to stepwise Frantz magnetic separation to remove the dense magnetic mineral fraction (e.g., pyrite and hornblende).

Separates were mounted in epoxy with several standards of Sri Lanka (known age 563.5 ± 3.2 Ma). Sample mounts were polished, imaged using reflected light and BSE (backscatter electron microscopy), and cleaned before analyses.

Analysis

Detrital zircon samples were analyzed for U and Pb isotopes using the LA-MC-ICPMS (laser ablation–multi-collector–inductively coupled plasma mass spectrometer) at the Arizona LaserChron facility at University of Arizona in Tucson (using methods outlined in Gehrels et al., 2008; Gehrels, 2012). Analyses were conducted with a Photon Machines Analyte G2 excimer laser. Analyses consist of single 15-second integrations on peaks with the laser off (for backgrounds), 15 one-second integrations with the laser firing, and a 30-second delay to purge the previous sample. The resultant ablation pits are ~15 microns in depth and 30 microns in diameter.

In-run analysis of Sri Lanka standards were conducted after every fifth unknown grain analysis. Data reductions for sample runs are performed in Isoplot program (Ludwig, 2008). Ages are corrected for $^{206}Pb/^{238}U$ and $^{206}Pb/^{207}Pb$ instrumental fractionation using fractionation factors determined from standards analyzed throughout the sample run using methods described in Gehrels et al. (2008) and Gehrels (2012) (see Table DR3 for analysis).

RESULTS

Stratigraphy

Two sections of the upper member of the Crystal Spring Formation were measured and described in detail (see Tables DR1, DR2; Figs. DR1, DR2). Sections were measured at Beck Canyon in the Kingston Range (base 11S_0597602, 3960080; UTM Zone 11, North American Datum 27) in the vicinity of the original type localities for the Pahrump Group (Hewett, 1940), and at Saratoga Spring in the southern Ibex Hills (base 11S_0552757, 3949353). These sections were chosen primarily to show the differences in the manifestation of the basal unconformity within the Crystal Spring Formation as well as to facilitate the assignment of type and reference sections to newly revised stratigraphic units as described below.

The stratigraphic sections from Beck Canyon in the Kingston Range and Saratoga Springs in the southern Ibex Hills represent the thickest and thinnest exposed sections of the upper member of the Crystal Spring Formation (Maud, 1979, 1983), respectively (see Tables DR1, DR2; Figs. DR1, DR2). The total thickness measured at Beck Canyon in the Kingston Range was 647.9 m along an approximately north-south transect. However, the location included several ~100-m-offset faults and as such, the lower portions (A and B units) of the section were measured ~700 m to the southeast of the remaining section (see Fig. DR3). The total thickness measured at Saratoga Spring is 132.6 m (see Fig. DR1), measured along an east-west transect. This section was structurally uninterrupted for the entire length of the measured section (see Fig. DR5).

Several stratigraphic trends are observable. Both sections show a dominance of clastic strata at the base fining upward through fine clastic strata and culminating in carbonate strata. The cyclic units of Maud (1979, 1983) are easy to correlate between sections, with the exception of the D and lower E units, which are absent in the Saratoga Springs area and indicate an intraformation unconformity (Figs. DR1 and DR2). Although this study did not focus on the facies analysis and paleoenvironmental interpretations of the upper Crystal Spring Formation, stratigraphic and sedimentologic observations made during this study are generally consistent with Maud's previous work and interpretations, and will be discussed in a later section.

A major unconformity is recognized between the middle and upper members of the Crystal Spring Formation, and separates two distinct stratigraphic sequences. This unconformity has been previously described (Maud, 1979, 1983; Mbuyi and Prave, 1993; MacDonald et al., 2013); however, little discussion was presented of the duration or changes in stratigraphic style across this unconformity. Here we describe this unconformity in greater detail and, in the following section, provide constraints on the magnitude of time represented.

Below the unconformity, sedimentary rocks of the middle member of the Crystal Spring Formation consist of thick (tens of meters) beds of predominantly coarsely crystalline stromatolitic dolomite, siltstone (commonly metamorphosed to hornfels), and silicified siltstone (described as the "cherty submember" by Maud [1979, 1983]). These sedimentary units are intruded by abundant, thick (several to tens of meters) diabase sills, and are commonly metamorphosed. Talc is commonly present where diabase sills intrude dolomitic rocks.

The unconformity is manifested by a slight angular discordance, from 0° to 20° (Mbuyi and Prave, 1993), and is overlain by coarse-grained siliciclastic units (coarse sandstone to cobble conglomerate). In the southwestern sections, the basal units commonly weather to dark brown to black, and contain angular cobbles of silicified or hornfelsic siltstone. Where the unconformity cuts diabase (observed by the authors in the southern Ibex Hills), diabase clasts are occasionally found in the overlying conglomeratic units. Siliciclastic rocks above the unconformity, particularly in southwestern sections, preserve 0.5-m-scale channel scours, as well as mudcracks and raindrop impressions in fine-grained siltstone-mudstone interbeds. In eastern sections (Kingston Range), the unconformity is more subtle and is overlain by coarse sandstone to pebble conglomerate. Clasts more commonly comprise silicified siltstone in a white to gray coarse sand matrix. Nowhere is diabase observed to intrude rocks above the unconformity.

Stratigraphic style above the unconformity is markedly different than below. A distinctive depositional cyclicity is manifested by siliciclastic rock units overlain by carbonate rock units, with cycles informally defined by Maud (1979, 1983) as A–F units. Cyclicity of this nature is notably absent below the unconformity. Strata above the unconformity are not contact metamorphosed as below and lack diabase sills, and stromatolite assemblages in upper member dolomites show significant age difference relative to lower- and middle-member assemblages (S.M. Awramik, 2012, personal commun.).

Geochronology

Of a total of 1098 detrital zircon grains analyzed from eight samples throughout the upper member of the Crystal Spring Formation using random sampling, a sample distribution (n = 6) of Neoproterozoic detrital zircon grains was found in four samples (see Fig. 4 and Tables 2 and 3 for samples collected containing Neoproterozoic grains). The remaining 1092 older grains will not be discussed further in this paper; these grains provide provenance information when combined with detrital zircon data from

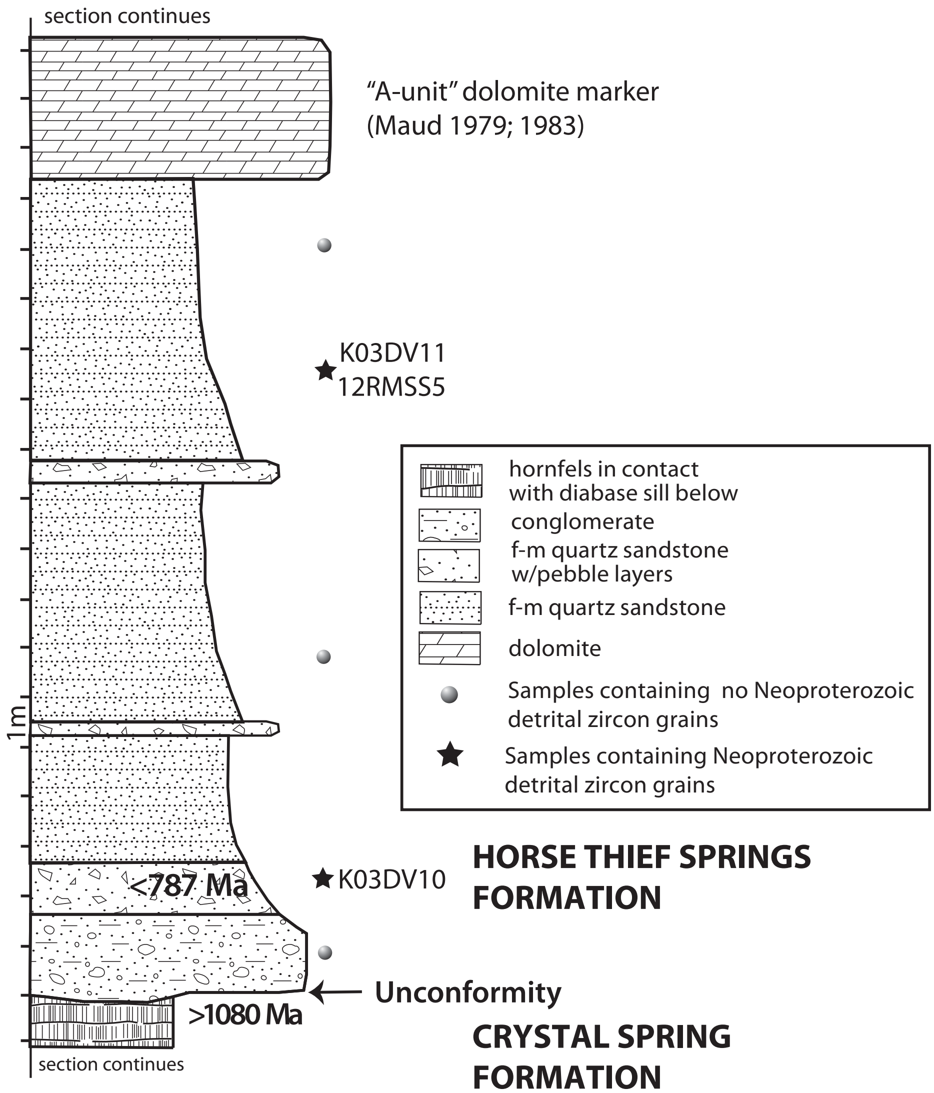

Figure 4. Detailed stratigraphic section of the base of the upper member of the Crystal Spring Formation from the southern Ibex Hills, showing locations where samples yielding young U-Pb detrital zircon grain ages were collected.

TABLE 3. U-Pb DETRITAL ZIRCON LA-ICP-MS ANALYTICAL RESULTS

Analysis	U (ppm)	$^{206}Pb/^{204}Pb$	U/Th	$^{207}Pb^*/^{235}U^*$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	Error corr.	$^{206}Pb^*/^{238}U$ age	± (Ma)	$^{207}Pb^*/^{235}U$ age	± (Ma)	$^{206}Pb^*/^{207}Pb$ age	± (Ma)	Best age (Ma)	± (Ma)	Conc (%)
K03DV10-21	57	6338	0.8	1.07361	3.0	0.12513	0.9	0.3	760.0	6.5	740.5	15.5	681.9	60.1	760.0	6.5	111.5
K03DV11-36	522	46,404	1.6	1.22891	1.8	0.13251	1.5	0.8	802.2	11.2	813.8	10.0	845.9	20.8	802.2	11.2	94.8

Analysis	U (ppm)	$^{206}Pb/^{204}Pb$	U/Th	$^{206}Pb^*/^{207}Pb^*$	± (%)	$^{207}Pb^*/^{235}U^*$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	Error corr.	$^{206}Pb^*/^{238}U$ age	± (Ma)	$^{207}Pb^*/^{235}U$ age	± (Ma)	$^{206}Pb^*/^{207}Pb$ age	± (Ma)	Best age (Ma)	± (Ma)	Conc (%)
K03DV10-6	1405	9092	6.1	13.8545	0.5	1.2913	2.6	0.1298	2.6	0.98	786.5	19.0	841.9	14.9	991.0	9.8	786.5	19.0	79.4
KO30V11-30	52	10,594	1.4	14.6941	7.1	1.2164	7.2	0.1296	1.5	0.21	785.8	11.2	808.1	40.2	870.2	146.3	785.8	11.2	90.3
12RMSS5-58	36	17,755	0.8	15.0539	3.6	1.1880	4.2	0.1297	2.3	0.54	786.2	16.9	795.0	23.3	819.9	74.2	786.2	16.9	95.9
4-CD11-30	62	15,753	1.1	15.4498	8.6	1.1394	8.7	0.1277	1.4	0.16	774.5	10.3	772.2	47.0	765.4	180.9	774.5	10.3	101.2

Note: See Table DR3 [see text footnote 1] for complete detrital zircon LA-ICP-MS (laser ablation–inductively coupled plasma–mass spectrometry) analysis notes. Corr.—corrected; Conc—Concordia; all errors reported ±1σ.

the remainder of the Proterozoic sedimentary rocks of the region, and will be the topic of a future manuscript. Two "young" grains were found in each of two samples (four grains total) overlying the unconformity between the middle and upper Crystal Spring Formation at Saratoga Spring (samples K03DV10 and K03DV11; see Table 3 for age data from each sample). One "young" grain each was found in samples overlying the aforementioned unconformity at Saratoga Spring (sample 12RMSS5) and from the F unit of the upper member of the Crystal Spring Formation at Crystal Spring in the Kingston Range (sample 4CD11). These young grains represent a significant population in that they are consistently present in multiple sample separates collected by multiple investigators, processed in different separation labs, and analyzed on multiple iterations of the same instrument over the course of nearly a decade. We therefore consider it implausible that these grains should represent contamination from other samples. Similarly, we reject the possibility that these grains yield young ages as a result of lead loss from older ca. 1000–1200 Ma grains, abundant in these samples, as this would be unlikely to result in such a narrow range of ages, given that no other grains analyzed were younger than 900 Ma. This would also result in reverse discordance in U-Pb and Pb-Pb ages, a result not observed with the grains presented herein (see Fig. 5).

A pooled weighted mean of ages from these grains (see Fig. 5) yields ages of 775 ± 18 Ma, with a mean square of weighted deviates value (MSWD) = 2.6, or of 787 ± 11 Ma, MSWD = 0.83, if one outlying data point is excluded (see Fig. 5). We take the latter age to be a more conservative estimate of the maximum depositional age. Five of the six grains were obtained from three samples from units directly overlying the unconformity at the base of the upper member of the Crystal Spring Formation (see Fig. 4); thus, this group of young grains lowers the maximum depositional age constraint for the basal part of the upper member of the Crystal Spring Formation to younger than 787 Ma. This age is some 300 m.y. younger than the ages reported from diabase sills in the lower and middle members of the Crystal Spring Formation (Heaman and Grotzinger, 1992) and constrains the duration of the unconformity between the middle and upper members of the Crystal Spring Formation to at least 300 m.y.

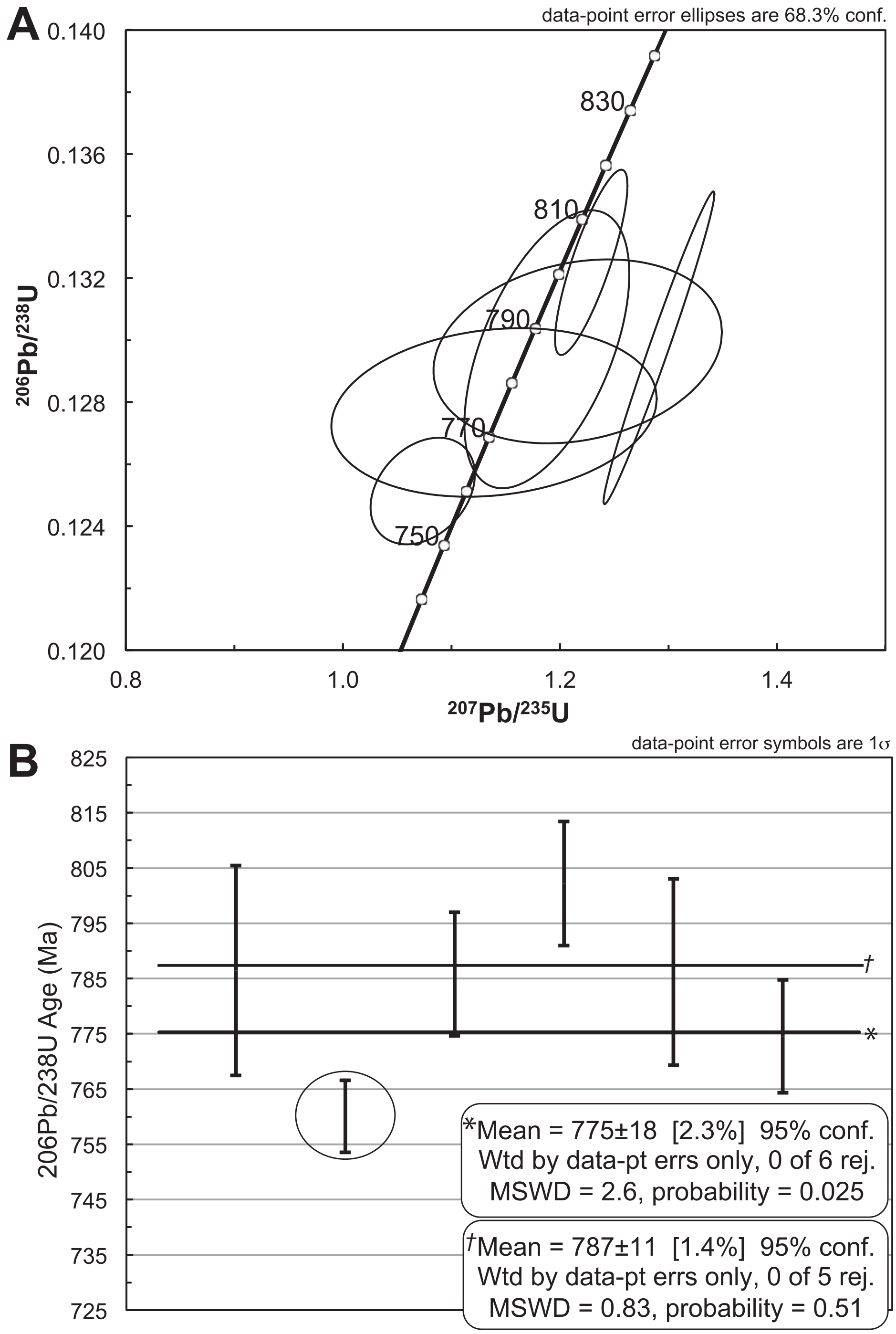

Figure 5. (A) Detrital zircon concordia plot showing ages of detrital zircon grains younger than ca. 810 Ma from samples K03DV10, K03DV11, 12RMSS5, and 4CD11 from the upper member of the Crystal Spring Formation (conf.—confidence). (B) Plot showing $^{206}Pb/^{238}U$ ages with calculated errors (with 1σ standard deviation) for six Neoproterozoic detrital zircon grains. A pooled weighted mean age of 775 ± 18 Ma with a mean square of weighted deviates (MSWD) value of 2.6 is calculated from grain age range of 760–802 Ma (n = 6). Presence of the youngest outlier results in a high MSWD value, therefore a separate pooled weighted mean age of 787 ± 11 Ma with MSWD value of 0.83 is calculated (n = 5) discounting the youngest age grain outlier (circled), providing a more conservative estimate for the maximum depositional age. Data shown in boxes are calculated means, errors and MSWD values at 95% confidence intervals for (*) all six grains and (†) five of six grains, rejecting the youngest outlier (circled). Abbreviations: Wtd—weighted; pt—point; errs—errors; rej—rejected.

STRATIGRAPHIC REVISIONS

Horse Thief Springs Formation

Detailed stratigraphic analysis of sections measured and described by Maud (1979, 1983) and those described by the authors, coupled with new detrital zircon age constraints, indicate that the unconformity at the base of the upper member of the Crystal Spring Formation represents at least 300 m.y. (Tables DR1, DR2; Figs. DR3, DR6). This unconformity separates two lithologically distinct units initially considered to be related (e.g., Hewett, 1940). These units were suspected by others to be unrelated to one another (Roberts, 1974a, 1974b, 1976, 1982; Maud, 1979, 1983; Mbuyi and Prave, 1993); however, with new detrital zircon age constraints, we are able to firmly quantify the temporal magnitude of this discrepancy and thus confirm the unrelated nature of these stratigraphic units. We propose to formally elevate the informal upper member of the Crystal Spring Formation (see Table 1; Fig. 2) above this unconformity (as described in detail above) to formation status and propose a name of Horse Thief Springs Formation. This name is derived from Horse Thief Springs (historically also referred to as Horse Spring; e.g., Hewett, 1956), located in the southeast corner of the U.S. Geological Survey (USGS) (1984) Horse Thief Springs 7.5′ quadrangle in the vicinity of the chosen type section described below.

A measured and described stratotype for the Horse Thief Springs Formation (see Tables DR1, DR2) is assigned at 11S_0597602E, 3960080N, east of Beck Spring in the western Kingston Range located on the USGS Horse Thief Springs 7.5′ quadrangle (U.S. Geological Survey, 1984) (Fig. DR3). This section is selected for the type section as it represents the thickest succession of the proposed formation (Maud, 1979, 1983) and is located in the vicinity of the original type localities and sections for each of the formations within the Pahrump Group (Hewett, 1940, 1956). This section, however, is variably exposed (Fig. DR4), and mapping in the area shows the area to exhibit some structural complexity (Calzia et al., 2000). Therefore, the type section measured spans several small drainages to accommodate faults and poor exposure (Fig. DR3).

Additionally, a reference section is assigned (see Table DR2) at 11S_0552634E, 3949335N, north of Saratoga Spring at the southern end of the Ibex Hills, located on the USGS Old Ibex Pass 7.5′ quadrangle (U.S. Geological Survey, 1985). The reference section is easily accessible from Saratoga Spring Road, less than 20 km from California Highway 127 (see Fig. 1). The area surrounding the reference section has been mapped at 1:24,000 scale and the section is known to be structurally uninterrupted at this locality (Figs. DR5, DR6), however this section represents the thinnest preserved sequence in the region (Maud, 1979, 1983). This section, while exceptionally well exposed, is not a complete representation of the Horse Thief Springs Formation, as the D unit is known to be absent in this location (Maud, 1979, 1983; Fig. DR2). It is selected as a reference section in order to provide an easily accessible location where the basal unconformity is exceptionally well preserved and exposed, and is the location from which detrital zircon grains were dated, yielding the age constraint presented above. Correlation between these sections illustrates facies variations from the eastern to western parts of the basin and the lateral traceability of several units within the new formation (see Maud [1979, 1983] for complete description of basin-wide correlations).

Analysis of cycles in the Horse Thief Springs Formation (described below) is consistent with observations presented in Maud (1979, 1983); unit divisions (A unit through F unit) developed therein were retained for the purposes of this study and are included in the stratigraphic summary of the Horse Thief Springs Formation presented in Table 4. We make no attempt to formalize these unit subdivisions as members for the purpose of this study, as they are considered "too thin to map at usual scales" (see North American Commission on Stratigraphic Nomenclature, 2005, p. 1560, Formal and Informal Units). Detailed stratigraphic descriptions for the type and reference sections are presented in Tables DR1 and DR2, respectively.

The upper interval in the Horse Thief Springs Formation (F unit of Maud, 1979, 1983), which exhibits the meter-scale, siliciclastic-dolomite cycles, gradually transitions into the overlying Beck Spring Dolomite. This contact is clearly definable in eastern exposures, where the overlying Beck Spring Dolomite is more monolithologic. Here we place the contact at the first appearance of gray to blue-gray laminated dolomite. However, this contact is gradational in western sections where the Beck Spring Dolomite includes a significant siliciclastic compo-

TABLE 4. STRATIGRAPHIC DEFINITION OF THE HORSE THIEF SPRINGS FORMATION

Unit	Description	Thickness (m)*
F unit		
	Siltstone, sandstone: red, green-gray. Interbedded dolomite: gray, orange. Dolomite interbeds more abundant in western sections. Dolomite microbially laminated. Orange, hematite-rich paleokarst with coarse siliciclastic infill on upper surfaces of some beds. Oncolitic chert locally. Upper contact transitional with Beck Spring Dolomite. Contact placed at first appearance of thick (>3 m) gray dolomite beds in eastern sections. In western sections, contact is more gradual, and placed where gray dolomite interbeds more abundant than varicolored dolomite.	50–80
E unit		
	Dolomite: gray, weathers brown. Stromatolitic, up to 15 cm diameter, 18 cm height. Silty dolomite between stromatolite mounds.	10–25
	Sandstone, some siltstone: green to gray. Bedded to massive. Arkosic. Local cross-beds. Locally present thin dolomite interbed. Oncolites present in chert beds: green.	3–40
D unit		
	Dolomite: light–medium gray, weathers brown. Microbial laminations. Intraclasts of carbonate common. Bedded chert.	0–40
	Siltstone, sandstone, quartzite, pebble conglomerate: green, tan-brown. Dolomite interbeds increasing upwards. Local trough cross-beds, oscillatory ripples.	0–100
C unit		
	Dolomite: medium gray, weathers brown. Stromatolitic. Interbedded siltstone: purple. Siltstone increases in proportion upward. Rare chert.	4–15
	Siltstone, sandstone, dolomite: light green to gray-tan, light red-brown. Thin bedded to laminated. Current ripple marks and trough cross-beds common in lower part.	10–100
B unit		
	Dolomite: gray to tan, weathers dark brown. Tabular to lenticular beds of chert. Microbial laminations, rare stromatolites.	1–10
	Siltstone and sandstone: light green-gray to brown. Thin bedded to laminated. Oscillatory ripples and flaser beds locally.	5–25
A unit		
	Dolomite: orange to tan. Microbial laminations, rare chert. Heavily recrystallized giant oolite (pisolite) in western sections. Sucrosic texture. Interbedded green-gray siltstone.	4–35
	Quartzite, conglomeratic: dark brown in lower part, white in upper part. Siltstone: purple to gray. Clasts predominantly dark green to black hornfels in lower part, and green, purple, red chert and siltstone in upper part. Quartzite exhibiting channel scour, trough and planar cross-bedding. Siltstone contains mudcracks and raindrop impressions in places. Locally ~20–40 cm of erosional scour at base of unit. Lower contact unconformable with underlying Crystal Spring Formation. Crystal Spring Formation below generally green hornfels or diabase, locally dark grey-green chert and silicified siltstone.	5–85

*Thickness ranges are generally presented from west (thinnest) to east (thickest).

nent and exhibits very similar meter-scale depositional cyclicity to the F unit of the Horse Thief Springs Formation. In these localities (namely in the Ibex and Saddle Peak Hills), the base of the Beck Spring Dolomite is difficult to determine. Consistent with the lithostratigraphic definition of the Beck Spring Dolomite (from Hewett, 1940), we place the boundary at the base of the first thick (>2 m), gray to blue-gray laminated dolomite unit. For the purpose of mapping and stratigraphic studies, this boundary is described as "transitional" in western sections, and as such, only approximately locatable.

Crystal Spring Formation

As a result of the proposed elevation of the upper member of the Crystal Spring Formation to the Horse Thief Springs Formation, a formal revision of the remainder of the Crystal Spring Formation is here defined. We propose to retain the name Crystal Spring Formation for the lower and middle members of the Crystal Spring Formation. The original lower boundary is retained (nonconformity between Mesoproterozoic metamorphic basement rocks and overlying sedimentary units; see Wright, 1952). A new upper boundary is defined by the unconformity separating the chert and hornfels of the middle Crystal Spring Formation and the basal conglomeratic and sandstone units of the Horse Thief Springs Formation (see Table 1; Fig. 2). This retention of nomenclature is permissible as the majority of the stratigraphic contents of Crystal Spring Formation are retained (~1000 m retained in eastern sections, with ≤640 m removed; ~800 m retained in western sections with ~150 m removed) and the proposed revisions result in significantly "more natural and useful" stratigraphic units [see North American Commission on Stratigraphic Nomenclature, 2005: Article 19, remark (a)]. No attempt is made herein to formalize any of the informal member subdivisions of the Crystal Spring Formation.

DISCUSSION

Stratigraphic Interpretations

Newly described sections and general paleoenvironmental interpretations largely agree with those from Maud (1979, 1983). In addition to prevalent mudcracks found in the finer-grained clastic intervals, enterolithic bedding (indicating deposition of evaporites, now removed by dissolution) is common, particularly in the A–C units, indicating deposition occurred in shallow, nearshore environments, perhaps within a restricted basin or sabkha-type environment (A.R. Prave, 2003, personal commun.; this study; Fig. 3E). The presence of symmetrical and asymmetrical ripple marks, trough cross-beds (Fig. 3F), and flaser beds (this study; Maud, 1979, 1983; Figs. DR1 and DR2) indicates that deposition throughout much of the Horse Thief Springs Formation occurred in current-influenced nearshore environments, above fair-weather wave base, such as in a barrier bar or open clastic shelf setting. Exposure features in the dolomitic facies (grikes, hematite-rich upper surfaces) are more prevalent than suggested by Maud (1979, 1983) and indicate that there were more episodes of subaerial exposure during carbonate deposition, especially in the upper part (F unit). Maud suggested several different models to explain the facies trends in the Horse Thief Springs Formation, and although the open clastic shelf model was preferred, it is more likely that a temporal and spatial spectrum of Maud's depositional models may provide the best understanding of the cyclicity and basin evolution. This task is beyond the scope of this paper and requires future detailed study.

The fining-upward and "cleaning"-upward trend that is apparent in the two sections measured, in combination with known sedimentology and cyclicity within the Horse Thief Springs Formation, suggests either a deepening-upward succession (from clastic to carbonate) and/or that the change in lithology may be controlled by the differential availability of coarser siliciclastic material (abundant in lower parts of sections) and coarser chemiclastic material (more dominant upsection). Similar trends are seen on the same scale (100 m scale) in the lithostratigraphic sequences of the correlative Chuar Group of Grand Canyon (see below). For example, the Kwagunt Formation fines upward and is overlain by shale interbedded with chemiclastic and stromatolitic dolomite facies associated with karst features (Dehler et al., 2001, 2005). The upper part of the correlative Uinta Mountain Group in Utah (see below) also shows a fining-upward succession from sandstone to shale (on a similar scale), yet no carbonate is present. Dehler et al. (2010) suggested that there was a regional or greater transgression during the deposition of the Uinta Mountain Group and Chuar Group, and the Horse Thief Springs Formation may be reflecting the same large-scale event. Dehler et al. (2005) called upon climate-controlled sea-level change to be responsible for cyclic lithologic changes (i.e., clastic to carbonate deposition) in the Chuar Group; this may also be the case for cyclicity in the Horse Thief Springs Formation.

Regardless of the specific depositional setting and stratigraphic patterns in the Horse Thief Springs Formation, it likely represents a marginal marine basin with a dominant source terrain to the south-southwest (Maud, 1979, 1983). Coarse clastic facies belts and paleoflow data were interpreted as defining a NW-SE–striking shoreline, with fining northward showing deepening to the north-northeast (Maud, 1979, 1983). However, these interpretations were made prior to tectonic reconstructions of the Death Valley region (e.g., Levy and Christie-Blick, 1991; Topping, 1993) and therefore the effects of Mesozoic contraction and Cenozoic extension were not considered. While lateral facies patterns and paleoflow data are seemingly consistent, regardless of location or transect orientation as reported by Maud (1979, 1983), shoreline trends, basin geometry, and paleocurrent results should be considered with caution. Further work is needed to fully characterize the facies trends of the Horse Thief Springs Formation in the context of more recent models for Mesozoic–Paleozoic tectonic reconstructions in the region.

Paleogeographic and Tectonic Implications

Detrital zircon maximum depositional age constraints from the newly named Horse Thief Springs Formation show that a ≥300 m.y. gap is represented by the basal unconformity, allowing more precise correlation with Proterozoic successions regionally (see Fig. 6). In particular, strikingly similar age relations and tectonostratigraphic packages are present in the Grand Canyon Supergroup which comprises both a ca. 1255–1100 Ma Mesoproterozoic (Stenian–Ectasian) sedimentary succession capped by 1100 Ma Cardenas Basalt (Elston, 1989; Timmons et al., 2005), and an unconformably overlying Neoproterozoic (Cryogenian) sedimentary succession of the 770–742 Ma Chuar Group (Karlstrom et al., 2000; Dehler et al., 2001, 2012). The Crystal Spring Formation and the Unkar Group indicate a regional tectonostratigraphic unit that formed in active rift basins responding to the Grenville orogeny during the assembly of Rodinia (Timmons et al., 2001, 2005). The Horse Thief Springs Formation, Beck Spring Dolomite, and likely "KP1" (*sensu* Prave, 1999; MacDonald et al., 2013) correlate with the Chuar Group, defining a regional tectonostratigraphic unit that formed in syn-extensional intracratonic basins during the early stages of the breakup of Rodinia (Timmons et al., 2001; Dehler et al., 2001, 2005).

This unconformity, therefore, is now known to be represented in stratigraphic successions across much of the western Laurentian margin. It is interpreted to be a major, regional feature associated with non-deposition and erosion during the period between amalgamation and rifting of the Rodinia supercontinent (e.g., Karl-

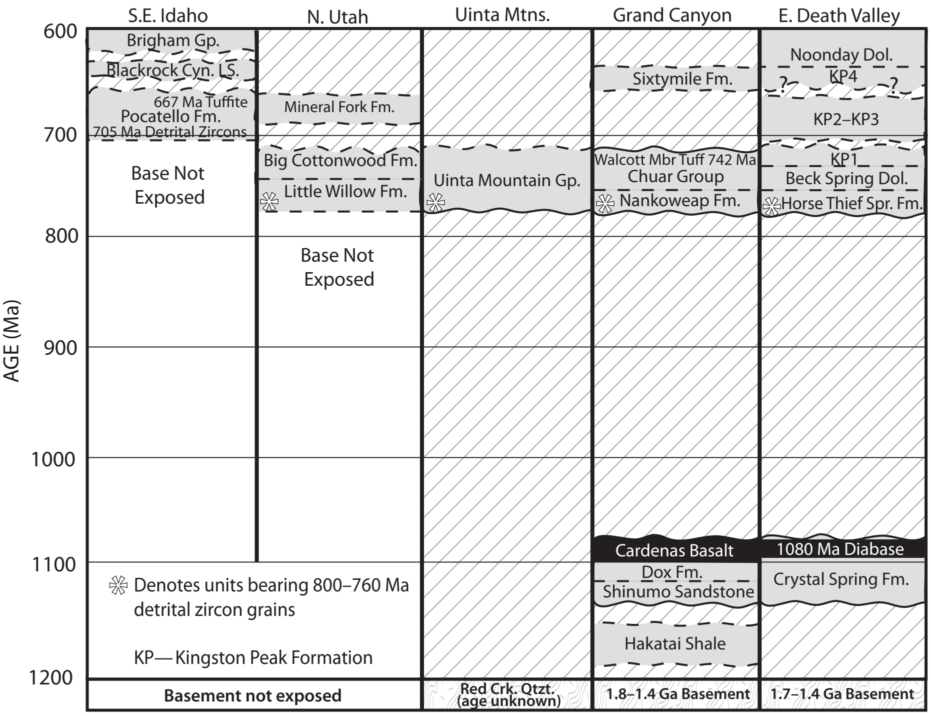

Figure 6. Refined correlation of Mesoproterozoic and Neoproterozoic sedimentary successions in southwestern Laurentia (modified after Link et al., 1993; Timmons et al., 2005; age constraints from Elston, 1989; Heaman and Grotzinger, 1992; Williams et al., 2003; Fanning and Link, 2004; Barth et al., 2009). Bold white asterisk indicates units containing 800–760 Ma detrital zircon grains (Timmons et al., 2005; Dehler et al., 2010, 2012; Spencer et al., 2012). Cyn.—Canyon; LS.—Limestone; Red Crk. Qtzt.—Red Creek Quartzite; Mbr—Member; Dol.—Dolomite; Spr.—Springs.

strom et al., 2000; Li et al., 2008). Detrital zircon grains of similar ages to the ones presented in this paper are known to be present in stratigraphic sequences immediately overlying this unconformity in the Grand Canyon and northern Utah regions marking a regional basin-forming episode beginning at ca. 780 Ma.

The revised Pahrump Group time-stratigraphic framework strengthens the use of vase-shaped microfossils (found in the Beck Spring Formation; Horodyski, 1993) as a reliable index fossil for ca. 780–740 Ma successions (Knoll, 2000; Porter and Knoll, 2000; Dehler et al., 2007), as well as brackets the timing of early advancements in food-web complexity of the global ecosystem (Karlstrom et al., 2000; Porter and Knoll, 2000; Nagy et al., 2009). Another result is the improved maximum age limit on the overlying diamictite units of the Kingston Peak Formation, interpreted to indicate glaciation (Miller, 1985), and possibly snowball Earth conditions (Hoffman et al., 1998) from ca. 720 to 635 Ma (Hoffman and Li, 2009; MacDonald et al., 2013).

The new age constraint on the Pahrump Group presented herein also allows for a refined mid-Neoproterozoic regional piercing point for western Laurentia that can be used to improve continental reconstructions of Rodinia and the geometry of the nascent Cordilleran passive margin. The Neoproterozoic grains discussed in this paper compose part of a larger set of detrital zircons sampled from the entirety of the Pahrump Group. Similar-age Neoproterozoic detrital grains have been found in several correlative stratigraphic units across the western North American margin (e.g., Dehler et al., 2010, 2012; Spencer et al., 2012) and no clearly defined magmatic source exists in western North America. Their widespread distribution in very small quantities is provocative, as it hints at the possibility that these grains may be derived from non–North American sources. This age population could therefore be present in proposed coeval Neoproterozoic basins on other continents and should be pursued. The implications of the provenance of these and other grain populations toward regional tectonics, paleocontinental reconstructions, and the development of the Proterozoic western Laurentian margin will be discussed in a future paper.

CONCLUSIONS

Six young detrital zircon grains found in a conglomerate and quartzite interval immediately overlying the unconformity between the lower–middle members of the Crystal Spring Formation and the upper member of the Crystal Spring Formation constrain the age of the upper Crystal Spring Formation to younger than 787 Ma. This new age constraint indicates that ≥300 m.y. is represented by an unconformity within the previously defined boundaries of the Crystal Spring Formation. The recognition of the magnitude of time represented by this unconformity, as well as the analysis of the stratigraphic dissimilarities between the strata above and below this unconformity, result in the elevation of the upper member of the Crystal Spring Formation to the Horse Thief Springs Formation and the revision of boundaries for the remainder of the Crystal Spring Formation. This geochronologic constraint and stratigraphic revision significantly clarify the regional correlations and the tectonostratigraphic history for Pahrump Group strata.

This regionally correlative unconformity, now recognized in strata across southwestern Laurentia, distinguishes three significant intervals in the development of the supercontinent Rodinia along the western Laurentian margin. The lower and middle Crystal Spring Formation represent inboard deposition during Mesoproterozoic (ca. 1250–1070 Ma) assembly of Rodinia. The unconformity at the boundary between the Crystal Spring Formation and the Horse Thief Springs Formation, similar to the Unkar Group–Chuar Group unconformity in Grand Canyon, represents relative tectonic quiescence along the western Laurentian margin during ca. 1070–780 Ma; and the overlying stratigraphic succession beginning with the Horse Thief Springs Formation through the Beck Spring Dolomite, Kingston Peak Formation, and the overlying Noonday Dolomite and Johnnie Formation represents the initiation of rifting through final separation along the southwestern Laurentian margin. These new tectonic-stratigraphic relationships from the Pahrump Group also provide important constraints on future studies attempting to determine the likely conjugate continental margin to the southwestern Laurentian margin during and before this time.

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