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**Cover image:** Mount Dunfee, near Gold Point, Nevada

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# Ediacaran-Cambrian transition of the Southwestern USA—Field Trip of the North American Paleontological Convention, June 19–22, 2019

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## SCHEDULE AND BRIEF ITINERARY

Participants should plan to arrive in Riverside by June 18. We will depart from the University of California, Riverside campus the morning of June 19. We will drive 5 hours to the White-Inyo Mountains. We will spend the remainder of June 19 and most of June 20 visiting the upper Ediacaran through lower Cambrian succession of this area, including a rich assemblage of trace fossils and some of the earliest Laurentian biomineralized fossils. We will drive to Beatty, NV the evening of June 20. On the morning of June 21, we will visit upper Ediacaran strata and an Ediacaran-Cambrian boundary section of the Reed Dolomite and Deep Spring Formation at Mount Dunfee and see Ediacaran microbialites, tubular body fossils and some of the oldest complex trace fossils. We will spend midday at the nearby ghost town of Gold Point, NV, and have an opportunity to see the spectacular early Cambrian archaeocyathan reefs of the Poleta Formation at Stewart's Mill in the afternoon. We will depart Beatty the morning of June 22 and return to Riverside by way of Death Valley, where we will make brief stops to discuss the stratigraphy of the Death Valley region, Neogene lake deposits near Shoshone and Cryogenian diamictites and the Marinoan cap carbonate exposed in the Saddle Peak Hills and at Sperry Wash. We will be staying in motels in Big Pine, CA (June 19) and Beatty, NV (June 20–21); participant lodging for these days will be covered by field trip fees. Lunches on all four days (June 19–22) and breakfast on the third day of the trip (June 21) will be provided; participants will be responsible for all other meals. We will return to Riverside in the late afternoon on June 22, in time for participants to check in to conference lodging. See Figure 1 for a map of field trip destinations.

**Tuesday, 6/18**: Participants arrive in Riverside. Spend night in Riverside (participants are responsible for making their own lodging and meal arrangements).

**Day 1—Wednesday, 6/19**: Meet at UC Riverside campus and drive to White-Inyo Mountains. Walk section of the Ediacaran Reed Dolomite and Ediacaran–Cambrian Deep Spring Formation at Hines Ridge in the afternoon. Spend night in Big Pine, CA.

**Day 2—Thursday, 6/20**: Visit exposures of the Campito, Poleta and Harkless formations near Westgard Pass. Drive to Beatty, NV; spend night in Beatty.

**Day 3—Friday, 6/21**: Drive to Mount Dunfee, near Gold Point, NV in the morning. Visit a section of the Reed Dolomite and Deep Spring Formation, containing Ediacaran microbialites, tubular body fossils and trace fossils. Spend midday at Gold Point. Optional afternoon visit to the archaeocyathan reefs in the Poleta Formation at Stewart's Mill. Spend night in Beatty.

**Day 4—Saturday, 6/22**: Drive to Death Valley, with a brief stop to view the Bare Mountain section outside of Beatty from the road. Brief stops in Shoshone, to discuss Pliocene–Pleistocene lacustrine deposits of Death Valley, and near the Saddle Peak Hills to visit the Cryogenian diamictites of the Kingston Peak Formation and see the Marinoan Noonday Formation cap carbonate. Return to UC Riverside by the late afternoon or evening, in time for participants to check in to conference lodging.

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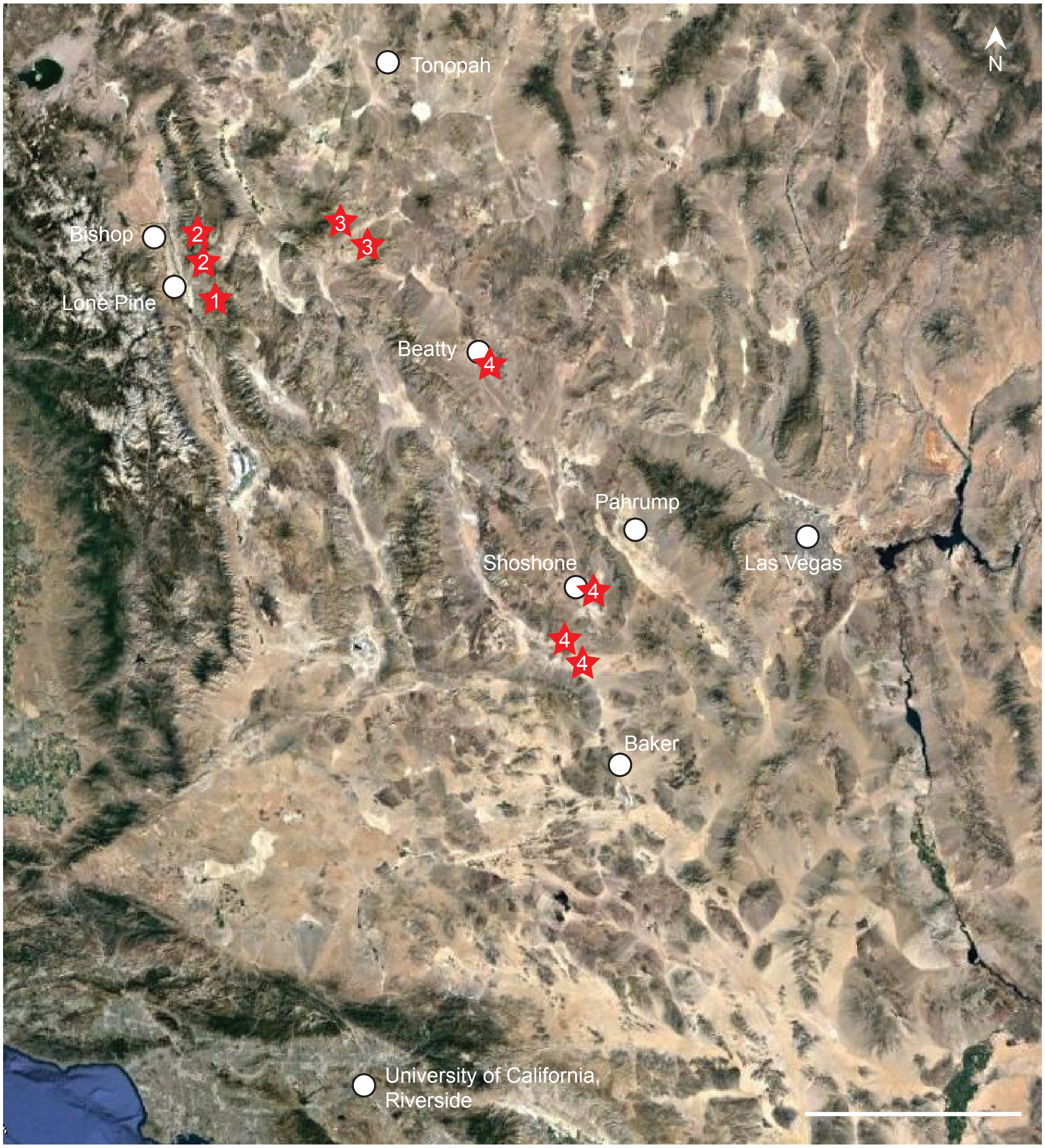



**Figure 1.** Regional satellite imagery of localities to be visited during this field trip. Numbers in stars denote day of the field trip during which that site will be visited. Scale bar=100 km.

## SAFETY INFORMATION

This field trip may involve hazards to the leaders and participants. Field trip participants are responsible for their own safety and wellbeing. We recommend sturdy hiking boots and long pants made of durable material, as well as sun protection.

The weather in the region, particularly in Esmeralda County and Death Valley, will be very hot. Participants should expect high daytime temperatures (>90˚ F), long hours of sun exposure and arid conditions. Nighttime temperatures, however, are likely to be considerably cooler (50–60˚ F). Participants are responsible for pacing themselves and not hiking past their personal limits. Please make sure you are prepared with adequate water and sun protection. Please be vigilant against dehydration; bear in mind that, in arid climates such as those we will encounter, it is easy (due to rapid evaporation of perspiration) to not be cognizant of the extent of fluid loss. We encourage all participants to bring water bottles. We will also provide jugs from which water bottles can be refilled, as well as bottled water. We encourage all participants to proactively hydrate and apply sun protection, and to familiarize themselves with the symptoms of dehydration and heat stroke. In the White-Inyo Mountains, we will be at high elevations. Please monitor yourself for signs of altitude sickness. If at any point you feel poorly, please communicate this to one of the field trip leaders or volunteers; please do not stay behind or seek to return to the vehicles without first notifying one of us.

Field trip participants are responsible for acting in a manner that is safe for themselves and their co-participants. This responsibility includes using personal protective equipment (PPE) when necessary, including safety glasses, gloves, and proper footwear. Participants should pay close attention to instructions from trip leaders at all field trip stops.

Additional hazards of this trip include rattlesnakes, traffic at highway stops, and mines. Rattlesnakes will certainly be out at this time of year, so please watch where you step, and step with intention. When climbing up outcrops, please be especially careful about where you place your hands; snakes like to hide in small crevices behind good handholds. There will be multiple stops along roads and highways. Please remember that your safety and the safety of others is more important than any rock or view. Finally, there are many mines in several of the areas we will visit—including partially collapsed and infilled pit mines which may have only a subtle surface expression. Participants are prohibited from entering any mine during this field trip. Please be aware of your surroundings and your footing at all times.

## HAMMERING AND COLLECTING POLICIES

Many of the sites that we will be visiting are classic field trip localities. They are frequently visited by professional geologists and paleontologists, as well as students. Some are also located on protected federal land. For this trip, we will have a no-hammering policy at all sites, in order to help preserve the rocks and fossils for future generations. If you would like to collect samples from non-federal land to take home, please collect them from float. Note that collection of any geological, paleontological, ecological or cultural samples from protected (e.g., US Forest Service) land is prohibited.

## INTRODUCTION

The Southwest USA has long been known as an extraordinary region for examining Ediacaran–Cambrian strata. Sections are easily accessible, well-exposed and well-studied. The Shuram negative carbon isotope excursion has been documented in the Johnnie Formation, which overlies the Marinoan cap carbonate and Cryogenian Snowball Earth glacial rocks. Late Ediacaran fossils have been documented from a number of localities and include specimens of classic Ediacara Biota taxa, a range of soft-bodied tubular organisms, *Cloudina* and diverse trace fossils. Additionally, the Ediacaran-Cambrian boundary has been documented, from the first appearance of the trace fossil *Treptichnus pedum* and the basal Cambrian negative carbon isotope excursion (known as the "BACE"), within sections of the Wood Canyon Formation and the correlative Deep Spring Formation. Lower Cambrian strata in this region consist of mixed carbonate-siliciclastic packages and include a number of potential biostratigraphic markers, including trace fossils, small shelly fossils, trilobites and archaeocyaths.

## GEOLOGIC BACKGROUND AND REGIONAL INFORMATION

### The Ediacaran-Cambrian transition in the Great Basin, USA

Upper Ediacaran through Cambrian strata in the southern Great Basin were deposited during the rift-to-drift transition along the western margin of Laurentia (Stewart 1970, Armin and Mayer 1983, Bond and Kominz 1984, Fedo and Cooper 2001). The classic view of Ediacaran–Cambrian basin architecture in the Great Basin is that it thickens from southeast to northwest, with siliciclastic-dominated successions in the Death Valley region deposited in a proximal shelf environment, and carbonate-dominated successions in the White-Inyo Mountains and Esmeralda County deposited in a more distal shelf environment (Fig. 2; Stewart 1970, Nelson 1976, Nelson 1978, Fedo and Cooper 1990, Mount et al. 1991, Corsetti and Hagadorn 2000, Fedo and Cooper 2001). Regional correlations and depositional basin reconstructions are hindered by the complex deformation history of this region, which has included multiple episodes of Mesozoic compression and subsequent Cenozoic extension (e.g., Wernicke et al. 1988, Snow et al. 1991, Snow and Wernicke 2000, Renik and Christie-Blick 2013, Pavlis et al. 2014). Although precise Ediacaran–Cambrian paleogeographic geometries remain uncertain due to these complexities, correlations between the Death Valley and White-Inyo Mountains strata are improving.

During this field trip, we will be visiting three regions in southwestern Nevada and southeastern California:

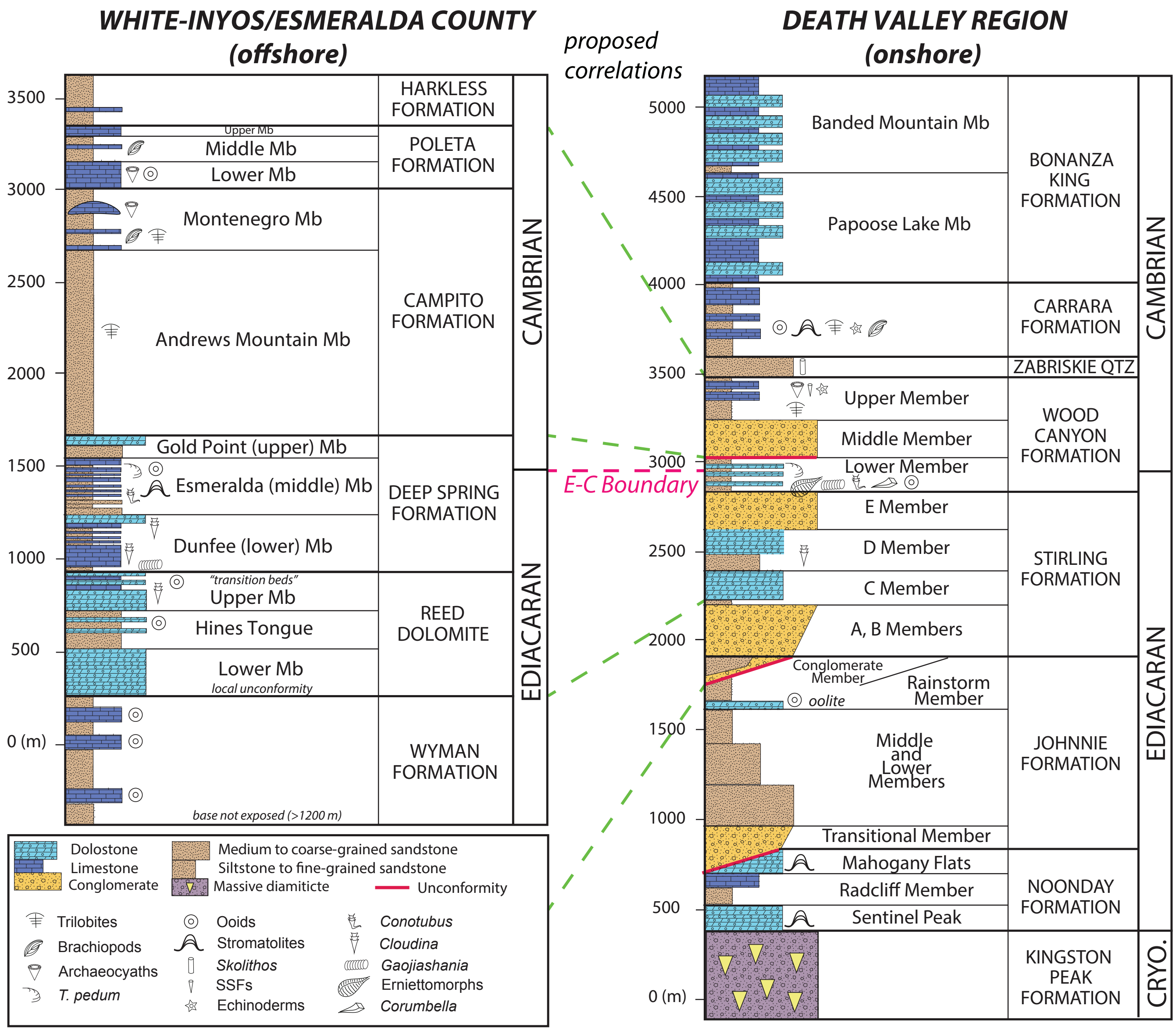



**Figure 2.** Schematic Ediacaran–Cambrian stratigraphic columns and proposed correlations between the White-Inyo Ranges/Esmeralda County and the Death Valley region.

1) the White-Inyo Mountains (Days 1–2), 2) Esmeralda County, NV (which hosts the same strata as the White-Inyo Mountains) (Day 3) and 3) the Death Valley region (Day 4). The trip will start in the more paleogeographically distal Ediacaran–Cambrian sections of the White-Inyo Mountains and Esmeralda County. On the final day of the field trip, we will briefly see the Cryogenian–Ediacaran succession of the southeastern Death Valley region, which preserves a phenomenal record of the Cryogenian Snowball Earth glacial episodes. Below, we summarize the Ediacaran–Cambrian and Tonian–Ediacaran stratigraphy of the southern Great Basin.

## Ediacaran–Cambrian stratigraphy in the White-Inyo Mountains and Esmeralda County

The oldest rocks exposed in the White-Inyo Mountains and Esmeralda County belong to the Wyman Formation, a unit which is, at localities in the White-Inyo Mountains, >1200 m thick, although Nelson (1962) reports an even greater minimum thickness of ~2700 m. The thickest continuous sections of the Wyman in Esmeralda County are ~400 m (Stewart 1970). It is composed dominantly of fine siliciclastic strata with interbeds of limestone to sandy limestone, which are commonly oolitic and are locally dolomitized.

The Wyman Formation is overlain by the Reed Dolomite. In the White-Inyo Mountains, an erosive unconformity has been locally documented at this contact (Nelson 1962, Lorentz 2007); however, it has also been argued to be a gradational contact (e.g., Albers and Stewart 1972). The Reed Dolomite is ~500 m thick and predominantly composed of massive featureless dolostone, but locally contains beds of ooids, pisoids, oncoids and microbialites (Stewart 1970). In the White-Inyo Mountains, the Hines Tongue is a laterally variable wedge of siliciclastic strata that occurs within the Reed Dolomite, ranging from 0 to 240 m in thickness between the lower and upper carbonate members (Nelson 1962). Elsewhere, however, the Hines Tongue is less discrete, and siliciclastic strata interbedded with carbonates occur throughout the middle to upper Reed Dolomite. The contact between the Reed Dolomite and the overlying Deep Spring Formation has been placed at a regionally distinct and recognizable sequence boundary, which is a sharp, karsted contact between underlying dolostone and overlying fine-grained siliciclastic facies.

In the White-Inyo Mountains and Esmeralda County, the Ediacaran-Cambrian boundary is in the Deep Spring Formation, which consists of the Dunfee, Esmeralda and Gold Point members (Corsetti and Kaufman 1994, Rowland and Corsetti 2002, Ahn et al. 2012). The Dunfee Member is 300–350 m thick and is composed primarily of grey to pale orange dolomitic limestone, with minor greenish-grey mudstone, siltstone, carbonate-cemented sandstone and quartzite (Albers and Stewart 1972, Gevirtzman and Mount 1986). Pink to blue recrystallized and locally karsted dolostone of the Dunfee Member is overlain by mud-cracked peritidal mudstone, interbedded with siltstone and sandstone, which is in turn overlain by tan to green shoreface and shelfal sandstone and siltstone. The overlying Esmeralda Member is ~250 m thick and composed of quartzite and calcareous sandstone, stromatolitic and ooilitic limestone and minor siltstone and shale (Albers and Stewart 1972). The Gold Point (upper) Member of the Deep Spring Formation sharply overlies the Esmeralda Member and consists of grayish-olive to greenish-gray siltstone and silty very fine-grained quartzite. The upper part of the Gold Point Member is a crystalline dolomite.

The late Ediacaran fossil *Cloudina* has been reported from the upper Reed Dolomite (Taylor 1966, Mount et al. 1983, Gevirtzman and Mount 1986, Signor et al. 1987, Grant 1990). Ediacaran soft-bodied fossils including *Gaojiashania*, *Conotubus*, *Wutubus*, and other unclassified smooth-walled tubular fossils have been found in two stratigraphic intervals in the Deep Spring Formation at Mount Dunfee (Smith et al. 2016b). A macroalgal fossil, *Elainabella*, has also been reported from the Esmeralda Member at Mount Dunfee (Rowland and Rodriguez 2014). Siliciclastic intervals within the lower Dunfee Member, as well as siliciclastic-dominated portions of the overlying Esmeralda and Gold Point members, contain trace fossils, which, locally, can be very abundant (Gevirtzman and Mount 1986, Smith et al. 2016b). The first appearance datum (FAD) of *Treptichnus pedum*, and thus the Ediacaran-Cambrian boundary (Landing 1994), has, at Mount Dunfee, been identified above the nadir of a large negative $\delta^{13}C$ excursion that has been correlated to the BACE (Corsetti and Hagadorn 2003, Smith et al. 2016b). Outcrops of the Gold Point Member in the White-Inyo Mountains contain the oldest regional examples of arthropod-produced trace fossils (*Rusophycus* and *Diplichnites*); however, these traces are very rare and not well-preserved (Alpert 1976).

A sharp contact separates the top of the Gold Point Member from the overlying Campito Formation, and has been interpreted as the base of the Sauk Sequence (Fedo and Cooper 2001, Corsetti and Hagadorn 2003). The Campito Formation is divided into three members: the Andrews Mountain, Gold Coin and Montenegro members. The Andrews Mountain Member consists of interbedded fine-grained quartzite, sandstone and siltstone; the Gold Coin Member is lithologically similar to the underlying Andrews Mountain Member, but also contains distinctive calcareous and frequently bioclastic nodules; the upper Montenegro Member consists of sandy and shaly siltstone and also, in Esmeralda County, contains an interval of calcareous, bioclastic nodules (Hollingsworth 2011). In Esmeralda County, the earliest regional trilobites, of the genus *Fritzaspis*, occur in the lowermost Gold Coin Member, within a few meters of the boundary with the underlying Andrews Mountain Member (Hollingsworth 2011). Trilobites, brachiopods and hyoliths occur throughout the Gold Coin and Montenegro members (Nelson 1978, Signor and McMenamin 1988; Hollingsworth 2011). The boundary between the *Fritzaspis* and *Fallotaspis* trilobite biozones (which marks the base of the Laurentian Montezuman Stage) is in the Montenegro Member, as is the boundary between the *Fallotaspis* and *Esmeraldina rowei* biozones; the boundary between the *Esmeraldina rowei* and *Grandinasus patulus* biozones occurs shortly below the top of the Montenegro Member (Hollingsworth 2011).

The Poleta Formation overlies the Campito and is generally ~360 m thick in Esmeralda County (though this varies widely) (Stewart 1970, Mount and Signor 1985). The Poleta Formation is characterized by a variety of lithologic types and an abundance of archaeocyathids and trilobites (e.g., Rowland et al. 2008, Hollingsworth 2011, Cordie et al. 2019). It is particularly well known for the massive archaeocyathan-microbial reefs in its lower member (Rowland 1984). The lower member is composed of shale, micritic ribbon limestone, thrombolitic limestone, grainstone, oolite and archaeocyath-microbial buildups (Rowland 1984, Rowland et al. 2008). It also contains small shelly fossils, such as *Lapworthella*, *Bemella* and *Hyolithellus*; problematic fossils such as the sigmoidal tubular organism *Lathamoserpens* (Waggoner and Hagadorn 2005); as well as inarticulate brachiopods (Gangloff 1975). The lower member contains the boundary between the *Grandinasus patulus* and *Avefallotaspis maria* trilobite biozones, and the lower-middle member boundary marks the base of the *Nevadia addyensis* trilobite biozone (Hollingsworth 2011). The middle member of the Poleta is siliciclastic-dominated, consisting of interbedded shale, siltstone, fine-grained sandstone and bioclastic limestone, recording a storm-dominated, shallowing upward succession (e.g., Stewart 1970, Mount and Signor 1985). The bioclastic limestone beds contain orthothecid hyoliths and trilobite sclerites as well as, more rarely, obolellid brachiopods. The boundary between the *Nevadia addyensis* and *Nevadella eucharis* trilobite biozones, as well as the boundary between the *Nevadella eucharis* and *Bonnia-Olenellus* trilobite biozones are placed within this member, with the latter also marking the boundary between the Laurentian Montezuman and Dyeran Stages (Hollingsworth 2011). The carbonate-dominated upper member, although less fossiliferous than the lower two members, contains trilobites, archaeocyaths and abundant trace fossils.

### Upper Ediacaran–Cambrian stratigraphy in the Death Valley region

In the Death Valley region, the upper Ediacaran Stirling Quartzite is gradationally overlain by the lower member of the Wood Canyon Formation. The contact is defined by a transition from massive white to grey cross-bedded mature quartz arenite to dominantly micaceous siltstone to fine sandstone with channelized beds of medium to coarse quartz sandstone over a few meters. The lower member of the Wood Canyon Formation consists of three parasequences of siltstone to sandstone, deposited in shallow subtidal environments, each capped by tan to buff dolostone marker beds, which are traceable across the Death Valley region (Stewart 1970, Diehl 1974). Simple bed-parallel trace fossils have been reported from siltstone and sandstone in the basal part of the lower member of the Wood Canyon Formation, as well as classic Ediacara Biota body fossils, such as erniettomorphs, and a variety of tubular body fossils, including *Conotubus* and *Gaojiashania* (Hagadorn and Waggoner 2000, Smith et al. 2017). The Ediacaran-Cambrian boundary in Death Valley is above the second dolostone marker bed of the lower member of the Wood Canyon Formation, where the FAD of *Treptichnus pedum* occurs just above the nadir of a broad negative carbon isotope excursion correlated to the BACE (Corsetti and Hagadorn 2000).

### Regional correlations between the Ediacaran–Cambrian of the White-Inyo Mountains/Esmeralda County and Death Valley

Although the White-Inyo Mountains and Esmeralda County are geographically close to the Death Valley area and contain units spanning broadly the same chronostratigraphic interval, precise regional correlations have been hindered by a lack of radiometric ages, sparse biostratigraphic control and regional structural complexity. Stewart (1970) proposed the first comprehensive lithostratigraphic correlation framework for the southern Great Basin. More recently, these correlations have been updated on the basis of carbon isotope and fossil occurrence data, suggesting that the Reed Dolomite and Dunfee Member of the Deep Spring Formation correlate with the Stirling Quartzite, and that the Wyman-Reed contact correlates with the contact between the upper Johnnie Formation and the Stirling Quartzite (Corsetti and Hagadorn 2000). In this framework, the Hines Tongue of the Reed Dolomite correlates with the lower to middle Stirling Quartzite, and the Wyman Formation is correlative to the Johnnie Formation (Corsetti and Hagadorn 2000). The lower member of the Wood Canyon Formation has been more precisely correlated to the Esmeralda Member of the Deep Spring Formation, on the basis of the FAD of *Treptichnus pedum* and an overlying negative carbon isotope excursion (interpreted as the BACE) (Corsetti and Kaufman 1994, Corsetti and Hagadorn 2000, Corsetti and Hagadorn 2003). The sequence boundary at the base of the middle member of the Wood Canyon Formation may be correlative with the base of the Campito Formation, and the Gold Coin Member of the Campito (which contains the FAD of *Fritzaspis* and presumably the base of the *Fritzaspis* biozone) may correlate to the lower part of the upper member of the Wood Canyon Formation (which also contains trilobite sclerites) (Stewart 1970, Hollingsworth 2011). The overlying

Harkless and Saline Valley formations in the White-Inyo Mountains and Esmeralda County are also, on the basis of biostratigraphic and lithostratigraphic comparisons, thought to correlate to the Zabriskie Quartzite and the lower members of the Carrara Formation of the Death Valley region (Stewart 1970). A summary of suggested correlations is presented in Figure 2.

### Tonian–Ediacaran stratigraphy in the Death Valley region

The Death Valley region of southeastern California hosts a 1.5–6.0 km-thick, well-exposed Mesoproterozoic to Cryogenian succession, the Pahrump Group, which is comprised of four formations overlying the basement and underlying the Noonday Formation, which has been correlated globally with the Marinoan cap carbonate. These formations are: the Mesoproterozoic Crystal Spring Formation, Tonian Horse Thief Springs Formation, Tonian Beck Spring Dolomite and Tonian–Cryogenian Kingston Peak Formation. Originally, it was thought that the units composing the Pahrump Group were deposited in a long-lived, failed aulacogen that was bound to the north and south by upland source areas, with carbonate-dominated units deposited during periods of tectonic quiescence (Wright et al. 1976, Roberts 1982). Over the past few decades, age constraints have been refined and several major unconformities have been recognized within the Pahrump Group, suggesting that Proterozoic strata in Death Valley were deposited during multiple distinct basin-forming events, or tectonostratigraphic units (TU), during the protracted breakup of Rodinia (Heaman and Grotzinger 1992, Prave 1999, Macdonald et al. 2013, Mahon et al. 2014, Smith et al. 2016a).

On this trip, we will briefly see the Kingston Peak and Noonday formations. In southern Death Valley, the Kingston Peak Formation has been divided into multiple informally named subunits. The lowest of these (KP1) consists of siltstone with no evidence of glacial activity and is interpreted as a Tonian pre-glacial unit. The next three subunits (KP2, KP3 and KP4) are interpreted as glacially derived diamictites, megabreccias and conglomerates recording the Cryogenian Snowball Earth glacial events and coeval tectonism (e.g., Prave 1999, Macdonald et al. 2013). In the Panamint Range, in western Death Valley, the Kingston Peak Formation clearly records both the Sturtian and the Marinoan glacial diamictites and cap carbonates (Prave 1999).

The Pahrump Group is overlain by the Noonday Formation, the lowest member of which has been identified as the basal Ediacaran cap carbonate (Prave 1999, Petterson et al. 2011). The Sentinel Peak Member of the Noonday Formation is composed of light-colored dolomicrite with irregular cements that locally contain tubestone stromatolite mounds (Cloud et al. 1974, Wright et al. 1978, Corsetti and Grotzinger 2005) and giant wave ripples (Creveling et al. 2016), distinctive sedimentary features of Marinoan cap dolostones globally (e.g., Allen and Hoffman 2005, Bosak et al. 2013). The Johnnie Formation disconformably overlies the Noonday Formation and is <1500 m thick, containing fluvio-deltaic siliciclastic rocks at its base that are overlain by marginal marine mixed siliciclastic and carbonate deposits (Stewart 1970, Summa 1993, Verdel et al. 2011). The upper, Rainstorm Member is the most widespread of the members of the Johnnie Formation. At the base of this member is a distinctive tan to orange oolitic bed (the "Johnnie Oolite") with a large negative $\delta^{13}C$ excursion that has been correlated globally with the Shuram carbon isotope excursion (e.g., Corsetti and Kaufman 2003, Halverson et al. 2005, Verdel et al. 2011).

The contact between the Johnnie Formation and the Stirling Quartzite is thought to be conformable in places (e.g., Stewart 1970), but others have interpreted it as a sequence boundary due to evidence for valley incision and an erosive unconformity within the Rainstorm Member (e.g., Christie-Blick et al. 1989, Summa 1993, Clapham and Corsetti 2005). The Stirling Quartzite (Nolan 1929) is composed primarily of quartz arenite, ranging from sandstone to pebble conglomerate, but also contains lesser siltstone and carbonate, particularly in more paleo-distal, northern and western sections (Stewart 1970). The Ediacaran mineralized tubular fossil (and potential index fossil) *Cloudina* has been reported from carbonate beds in the informal D member of the Stirling Quartzite in the Funeral Mountains (Langille 1974). The Stirling Quartzite is gradationally overlain by the Ediacaran–Cambrian Wood Canyon Formation, which is described above.

## FIELD TRIP STOPS

### Day 1 (6/19) morning: Departure from University of California, Riverside

*We will be departing from the University of California, Riverside campus at 8:00 am sharp. Please plan to arrive at the vans by 7:30 am. Participants are responsible for their own breakfasts and for transporting themselves to the UC Riverside campus. Please check in with a trip leader once you arrive at the rendezvous point.*

*UC Riverside will provide sack lunches for this first* *day; please pick up your sack lunch prior to boarding the vehicles. Please do not take one of the lunches marked "vegetarian/vegan" unless you specifically requested a vegetarian lunch. Participants should eat their lunches in the vehicles; we will not be making a formal lunch stop. We will be making a couple of brief stops to refuel the vehicles and use the restrooms, but will not reach our first field trip stop until the early afternoon.*

**Day 1 (6/19) afternoon: Ediacaran–Cambrian strata of the White-Inyo Mountains**

*We will be spending this afternoon at mid-range elevations (<2400 m or ~7800 ft), and the temperature will likely be very hot. We will start with a hike up rougher and somewhat steep topography, before reconvening for our first stop. Participants should hike at their own pace, and take care not to exceed their personal limits. Please be sure to stay hydrated and to protect yourself from sun exposure, and familiarize yourself with the symptoms of heat stress and altitude sickness (see enclosed handouts). Please inform a trip leader if you feel unwell; do not attempt to return to the vehicles without first notifying one of us.*

*Stop 1.1. Overview of Hines Ridge Ediacaran–Cambrian section*—Hines Ridge is the site of a relatively continuous succession of the Ediacaran–Cambrian units in the White-Inyo Ranges and Esmeralda County, NV. These units include, in successive order, the siliciclastic-dominated Ediacaran Wyman Formation, the Ediacaran Reed Dolomite, the Ediacaran–Cambrian Deep Spring Formation and the Cambrian Campito Formation. Younger Cambrian strata of this succession are also exposed in the surrounding hills and the slopes of Andrews Mountain. This afternoon we will see the Reed Dolomite and the Dunfee (lower) and Esmeralda (middle) members of the Deep Spring Formation—units which we will return to in Esmeralda County, NV, on Day 3 of the field trip. See Figure 3 for a map of Hines Ridge field trip stops.

*Stop 1.2. Ediacaran Reed Dolomite, crest of small knoll*—We are standing in the uppermost Reed Dolomite, just a few meters below the boundary with the Dunfee Member of the Deep Spring Formation. This contact is also clearly visible on the main ridge to the north (along which the remainder of today's field stops are located). In intervals of the upper Reed Dolomite in the White Mountains and in Esmeralda County, grainstone beds containing sclerites of cloudinid tubular fossils have been reported (e.g., Taylor 1966, Gevirtzman and Mount 1986, Smith et al. 2016b).

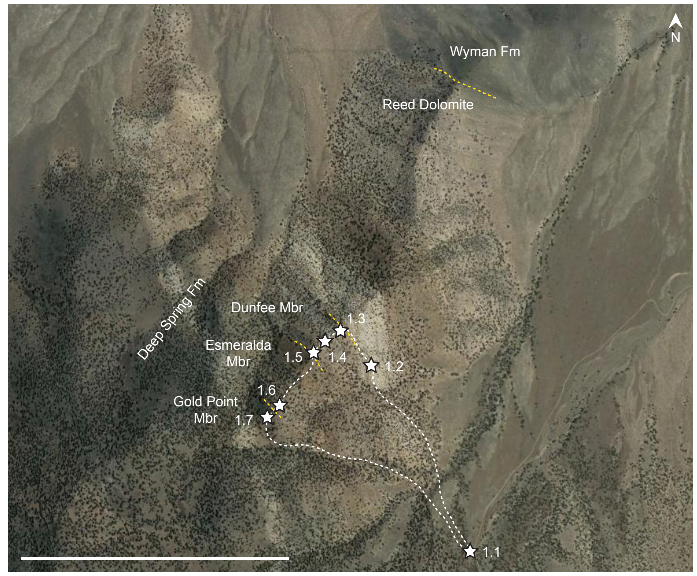



**Figure 3.** Satellite image of Hines Ridge, showing stops and walking path for the afternoon of Day 1 of the field trip. Scale bar=1 km.

*Stop 1.3. Boundary between Reed Dolomite and Dunfee (lower) Member of the Deep Spring Formation*—We have just crossed the boundary between the Reed Dolomite and the Dunfee Member of the Deep Spring Formation, both of which are upper Ediacaran in age. The grainstone carbonate beds on which we are currently standing are characteristic of both the upper Reed Dolomite and the Dunfee Member of the Deep Spring Formation. In the White-Inyo Mountains, this contact is defined by a change from a massive white recrystallized dolostone (the Reed Dolomite) to limestone interbedded with shale, siltstone and sandstone (the lowermost Deep Spring Formation). In Esmeralda County, this contact is defined by a clear, regionally persistent exposure surface (Smith et al. 2016b). Cloudinids have been reported from throughout the Dunfee (lower) Member of the Deep Spring Formation from localities in Esmeralda County (e.g., Smith et al. 2016b).

*Stop 1.4. Upper Dunfee (lower) Member of the Deep Spring Formation*—We are currently walking through facies characteristic of the upper Dunfee Member, including cross-bedded carbonate-cemented sandstone and decimeter-scale soft-sediment deformation. The prominent dolostone at the ridge crest above us marks the top of the Dunfee Member; the overlying cliff-forming sandstone marks the base of the Esmeralda Member. This sharp contact between the members is marked by an exposure surface. At localities in Esmeralda County (such as the section at Mount Dunfee that we will see on Day 3 of the field trip), grainstones containing cloudinid sclerites occur throughout the the Dunfee Member (Smith et al. 2016b).

*Stop 1.5. Lower Esmeralda (middle) Member of the Deep Spring Formation*—The lithology of the lower Esmeralda (middle) Member is siliciclastic dominated. Indications of emergent to very shallow marine conditions, such as mud cracks, poorly sorted mudstone intraclasts and interference ripples, are common. At this locality, trace fossils in this interval are relatively rare and morphologically simple trails and infilled burrows.

*Stop 1.6. Esmeralda (middle) Member stromatolites*—Due to recrystallization and foliation, the regionally extensive stromatolites of the Esmeralda (middle) Member are not as easy to see in this area as they are in other sections; at this stop, however, it is possible to see one of the several stromatolitic horizons in this unit. At Mount Dunfee, these have been described and studied in detail, and are part of various Ediacaran–Cambrian reefs that are collectively referred to as "Rowland's Reefs" (Oliver 1990, Oliver and Rowland 2002, Rowland et al. 2008). The $\delta^{13}C$ values of these reefs are isotopically negative (approximately -6 ‰), and they crop out just below the nadir of a large negative carbon isotope excursion that has been correlated globally with the BACE.

*Stop 1.7. Boundary between Esmeralda (middle) and Gold Point (upper) members of the Deep Spring Formation*—We have just crossed the boundary between the Esmeralda and Gold Point members of the Deep Spring Formation, marked by the top of a white to blue limestone. In Esmeralda County, the Ediacaran-Cambrian boundary—as marked by the FAD of the three-dimensional, sediment-penetrative burrow *Treptichnus pedum*—occurs within the upper Esmeralda Member. Lithostratigraphic correlation suggests that, at Hines Ridge, this interval should correlate with the siltstone-dominated package through which we just traversed. Treptichnid burrows have been found here in the sandstones at the top of this package, and both treptichnids and *Treptichnus pedum* can be seen in the Gold Point Member of the Deep Spring Formation. The overlying Andrews Mountain Member of the Campito Formation is also rich in trace fossils.

*Return to vehicles via wash to south of Stop 1.7. Drive from Hines Road back to Big Pine, CA. Check into hotels. We have a group reservation for dinner at the Country Kitchen, a short walk from both hotels. Participants are responsible for paying for their own dinners.*

## Day 2 (6/20): Lower Cambrian succession of the White-Inyo Mountains

*Breakfast in Big Pine; participants responsible for own breakfast; check out of hotel and depart for field by 8:30 am*

*We will spend all of today visiting localities in a heavily visited area of Inyo National Forest. Please be respectful of Forest Service rules and regulations; collection of any materials (e.g., geological, paleontological, ecological or cultural) is strictly prohibited. Please take care not to litter and please be respectful of other visitors to this area. Please be especially mindful of traffic along the road, as traffic may be relatively heavy and visibility is frequently limited.*

*We will be at higher elevations today (<2850 m or 9350 ft). Please take care to stay hydrated, protect yourself from sun exposure and heat stress and monitor yourself for symptoms of altitude sickness. The trip leaders will have pain medication for altitude-related headaches.*

*Stop 2.1. Sierra View Point: overview of Ediacaran–Cambrian stratigraphy of the White-Inyo Mountains*—Today we will visit outcrops characteristic of the Cambrian of the White-Inyo Mountains (Fig. 4), and continue to develop a stratigraphic framework for the Ediacaran–Cambrian of both this region and of neighboring Esmeralda County, NV, which we will visit on Day 3 of the field trip.

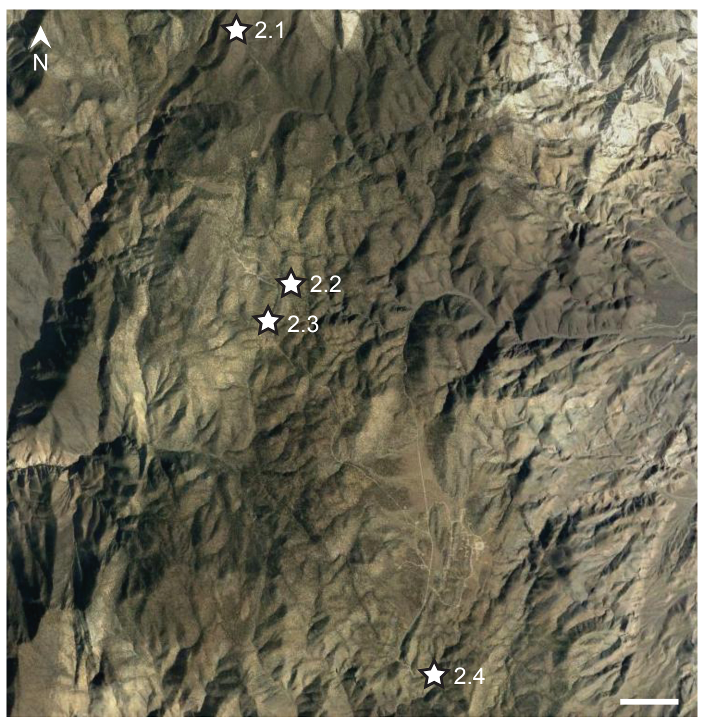



**Figure 4.** Satellite image of Westgard Pass region of White-Inyo Ranges, showing stops for Day 2 of the field trip. Scale bar=1 km.

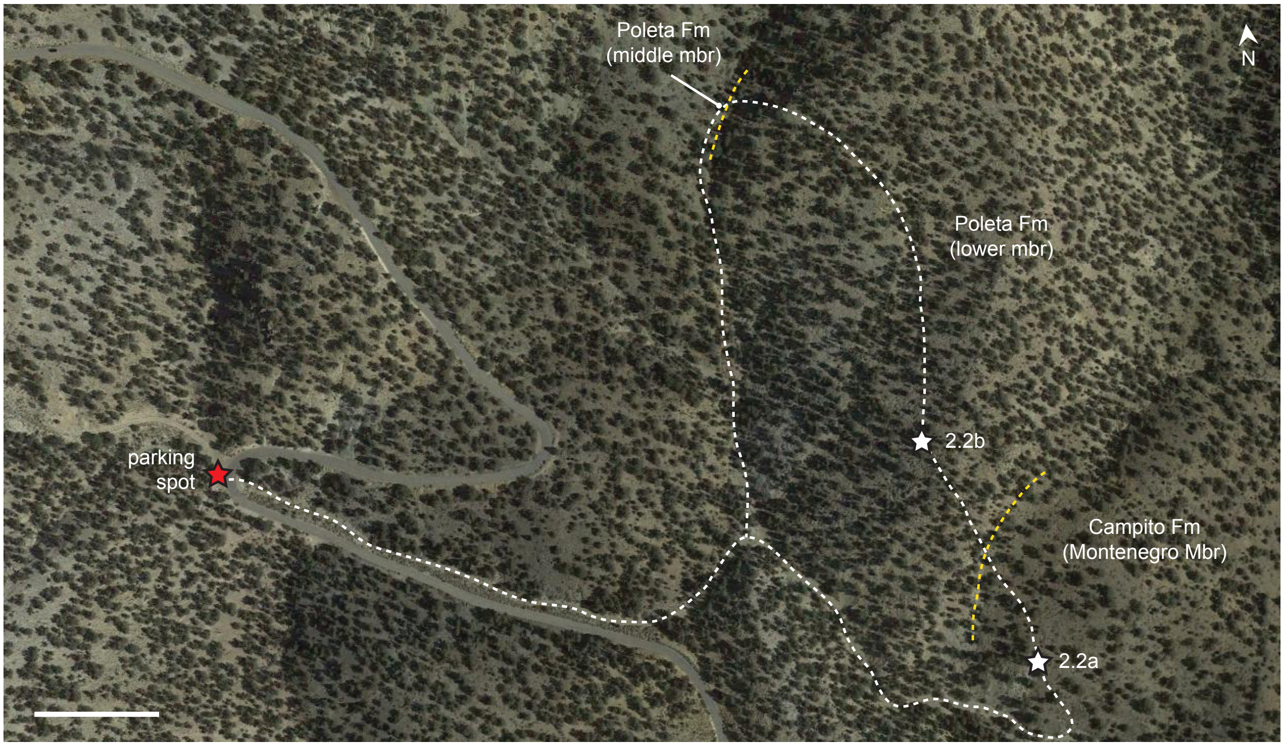



**Figure 5.** Satellite image of Stop 2.2 of the field trip, showing sub-localities and suggested walking path through the Montenegro Member of the Campito Formation and the lower Poleta Formation. Scale bar=100 m.

This morning and afternoon we will be making stops at various sites along California highway 168 and White Mountain Road in the vicinity of Westgard Pass (Fig. 4), which divides the White Mountains in the north from the Inyo Mountains in the south.

From this view point we can, looking southeast in the foreground, see the striped light-colored Ediacaran Reed Dolomite and Ediacaran–Cambrian Deep Spring Formation. To the south we can see hills of the Cambrian Campito and Poleta formations, which we will visit at several stops throughout the day. Looking south beyond CA-168, we can see the Inyo Mountains. The prominent peak to the southwest is Waucoba Mountain; the smaller peak slightly north of Waucoba Mountain is Andrews Mountain and the Hines Road locality we visited yesterday afternoon (Stops 1.1–1.7). To the east, across the California-Nevada border, is Esmeralda County, which we will visit tomorrow. To the west is the Owens River Valley (including the town of Bishop, CA), which has been prominently shaped by both glacial and volcanic activity. West of the Owens River Valley are the Sierra Nevada.

The rocks on which we are currently standing are part of the lower Cambrian Andrews Mountain (lower) Member of the Campito Formation. Trackways and trails are common in the Andrews Mountain Member (Nelson and Durham 1966).

*Stop 2.2. Montenegro and Poleta archaeocyathid bioherms*—We will park on the northwest side of the road. Please CAREFULLY cross to the southeast side of the road and enter the wash running parallel to the road. Follow this wash downhill and to the left, as it steepens to a shallow canyon. The cliffy 'narrows' of this portion of the canyon are composed of the lower member of the Poleta Formation. Follow the wash to the right; we are walking stratigraphically down-section through the heterolithic upper interval of the Montenegro (upper) Member of the Campito Formation. Our first stop will be at the next carbonate-dominated interval. See Figure 5 for a map of field trip stops at this locality, and a suggested walking route.

The Montenegro Member locally marks the base of the Montezuman Stage of Laurentia. Trilobites of the *Fallotaspis* biozone, including *Montezumaspis* cf. *cometes* (specimens originally described by Nelson and Hupé 1964; subsequently taxonomically revised to *Montezumaspis* cf. *cometes* (e.g., Fritz 1995; Hollingsworth 2005, Hollingsworth 2006, Hollingsworth 2007)) and *Cirquella nelsoni* and/or *Daguinaspis* sp. occur in the Montenegro Member of the Westgard Pass area (Hollingsworth, 2011). In Esmeralda County, the Montenegro Member also includes the overlying *Esmeraldina roweii* biozone and the base of the *Grandinasus patulus* biozone (Hollingsworth 2007, Hollingsworth 2011).

*2.2a: Montenegro archaeocyathid bioherm*—The upper Montenegro Member hosts bioherms containing a distinctive archaeocyathid assemblage. These buildups, which in places have recorded thicknesses of >5 meters, contain members of the genera *Ethmophyllum* and *Archaeocyathus*, which have columnar, branching morphologies and are inferred to have had a solitary life mode (Nelson and Durham 1966). These buildups also include echinoderm and trilobite debris (Gangloff 1976), as well as a substantial microbial and 'abiogenic' component (Kiessling 2002, Cordie et al. 2019). Although archaeocyaths may have contributed to framework-building, the Montenegro bioherms in the White-Inyo region consist primarily of micritic matrix, with minor contributions from *Renalcis*-type microbial encrusters (e.g., Xiaoping 1995, Cordie et al. 2019). At this locality, biohermal buildups in the upper Montenegro Member occur discontinuously along strike, with smaller bioherms occurring intermittently in the overlying strata of the uppermost Montenegro Member (Nelson and Durham 1966).

*2.2b: Lower Poleta archaeocyathid bioherms*—As we move up-section into siltstones and sandstones, we will have crossed the boundary between the Montenegro (upper) Member of the Campito Formation and the lower member of the Poleta Formation. As we walk through the lower Poleta Formation, we will see that the dominant lithology consists of thin- to medium-bedded blue-gray limestone, with local archaeocyathid bioherms. The lower Poleta bioherms commonly contain members of the genus *Archaeocyathus* (Nelson and Durham 1966). Recent analyses (Cordie et al. 2019) have (in agreement with prior work; e.g. Rowland and Gangloff (1988)) indicated that, although the metazoan contribution to framework construction in the Poleta Formation (up to 15% in individual samples) is greater than in the underlying Montenegro bioherms, archaeocyaths in these bioherms are patchily distributed and occur largely in cavities (Cordie et al. 2019). Micrite and calcifying microbes, such as *Renalcis*, still constitute the major components of Poleta bioherms (Cordie et al. 2019). Poleta Formation bioherms are characterized by higher richness and lower evenness values than Montenegro Member bioherms, suggesting that the buildups of the lower Poleta Formation record more diverse but also more sparse and heterogeneous ecosystems, relative to those of the Montenegro Member (Cordie et al. 2019). Unlike the bioherms of the Montenegro Member, archaeocyathid bioherms in the lower Poleta Formation contain colonial as well as solitary forms (Nelson and Durham 1966). Archaeocyathids in this interval have been compared to those of the lower Cambrian of British Columbia and the Yukon, Canada (Gangloff 1976). Grainstone, locally dominated by echinoderm plates, and oolitic carbonate are also common in this interval; some bioherms are capped by oolite beds, suggesting that these organisms lived in relatively high-energy shallow marine settings, such as reef and ooid shoal complexes (Rowland and Shapiro 2002). Blue-gold carbonates characterized by low to moderate intensities of bioturbation are also common throughout this succession. Trace fossil assemblages include horizontal forms, *Thalassinoides*-type branching burrows, and vertical J-shaped and *Arenicolites*-type U-shaped burrows. The top of the lower member of the Poleta Formation is sharply overlain by the siltstones and shales of the lowermost middle member.

Upon reaching the contact between the lower and middle members of the Poleta, we will walk down the narrow wash, rejoining the main canyon. Turn left to return to the confluence between this canyon and the narrow canyon leading up to White Mountain Road; when reaching the point along the road from which we originally entered, please cross the road cautiously to return to the vehicles.

*Stop 2.3. Pinyon Nature Trail—We will have lunch at this stop. At this locality, please do not stray from the designated trail.*

At this stop, a half-mile trail provides the opportunity to see additional examples of archaeocyathid bioherms in the lower member of the Poleta Formation. Colonial archaeocyathids are visible in some of the outcrop exposures along this trail. Blue-gold moderately bioturbated carbonates and carbonate oolite beds are also common in this interval. Near the end of the trail (if a counterclockwise loop is taken) we can once again see the boundary between the oolitic and bioturbated carbonates of the uppermost lower Poleta and the heterolithic siliciclastics of the lowermost middle Poleta.

Tomorrow afternoon we will see additional examples of lower Poleta archaeocyathid bioherms at Stewart's Mill near the town of Gold Point, NV (Stop 3.5).

*Stop 2.4. Middle–upper Poleta and lower Harkless formations—We will park on the western side of the highway. This stop is located in a narrow canyon on the eastern side of the next bend to the southeast in the highway. Please* *cross the road, with caution, at our parking spot (do not attempt to cross at another point) and immediately enter the small wash running parallel to the road. Follow this wash until reaching the narrow canyon that marks the beginning of this stop.*

Here we will see exposures of the middle–upper Poleta and the lower Harkless formations (Fig. 6). The middle Poleta Formation in this region is heterolithic, consisting of interbedded quartzite, grey-green siltstone and shale, thin and frequently bioclastic cream-colored and light blue limestone beds and burrowed blue-gold limestone (Hollingsworth 2011). The dark red thick quartzite ledge in this section contains "piperock" of densely packed vertical *Skolithos* burrows. The upper Poleta Formation is dominated by thick-bedded blue-gray carbonate that locally contains archaeocyathids. The upper member of the Poleta Formation is sharply overlain by the Harkless Formation, which consists of gray-green shale with pisolitic carbonate interbeds (Nelson and Durham 1966). Detailed characterization of both horizontal and vertical bioturbation in siliciclastic intervals of the Campito, Poleta and Harkless formations in the White-Inyo Mountains (including this locality) has indicated that the intensity of vertical bioturbation is, apart from thin and sparse intervals of piperock, extremely limited in these successions (with Ichnofabric Index (II) values, (cf. Droser and Bottjer 1986; II 1 = laminated, II 6 = homogenized) ranging from II 1 to II 2) (Droser 1987, Marenco and Bottjer 2008). Horizontal burrowing, however, is considerably more common, and bedding-plane trace fossil assemblages are typically dominated by the unlined horizontal burrow *Planolites*, with relatively rare occurrences of arthropod-produced *Rusophycus*, *Cruziana* and isolated scratch marks and the potential actinian resting burrow *Bergaueria*. Bedding-plane surfaces in the siliciclastic-dominated intervals of this succession are moderately well-bioturbated (with Bedding Plane Bioturbation Index (BPBI; cf. Miller and Smail 1997) values ranging between BPBI 1 (0% bioturbation) and BPBI 5 (100% bioturbation) (Marenco and Bottjer 2008). Carbonate intervals of the White-Inyo Mountain Cambrian succession typically range from poorly bioturbated to moderately well bioturbated (e.g., Droser 1987, Droser and Bottjer 1988, Tarhan 2018).

*Return to vehicles via small wash on northeast side of highway (cross road, with caution, only when adjacent*


**Figure 6.** Satellite image of Stop 2.4 of the field trip, with suggested walking path through the middle and upper members of the Poleta Formation. Scale bar=100 m.

*to vehicles). Drive from White-Inyo Mountains to Beatty, Nevada. Check into hotels. We have a group reservation for dinner at the Sourdough Saloon, a short (0.6 km) walk from the hotels. Participants are responsible for paying for their own dinners.*

**Day 3 (6/21) morning: Ediacaran strata of Esmeralda County**

*Depart hotels no later than 6:00 am. There will be coffee and breakfast, provided by field trip organizers, as well as restrooms, once we arrive at Gold Point.*

*We will be in lower-elevation terrain and thus experience hotter temperatures today. Please take care to stay hydrated and be mindful of sun exposure; do not exceed your personal hiking limits.*

One of the reasons why this field trip (and many other geology field trips and classes) is visiting the Ediacaran–Cambrian succession of Esmeralda County is that the strata here are of lower metamorphic grade than the correlative strata in the White-Inyo Mountains to the west. However, superimposed Mesozoic contractional and Cenozoic extensional faulting, lateral facies changes, and variable dolomitization of the upper Reed Dolomite and the lower part of the Deep Spring Formation have, historically, hampered attempts to continuously piece together the composite stratigraphy of Ediacaran and Cambrian units in this region. Recent detailed geologic mapping (Fig. 7) of the Mount Dunfee area, in particular, has allowed for a more complete stratigraphic study of Ediacaran–Cambrian strata and resulted in the discovery of horizons of Ediacaran body and trace fossils (Smith et al. 2016b). Today we will be visiting Ediacaran–Cambrian strata at Mount Dunfee and Stewart's Mill (Fig. 8).

*Note: A number of research projects (including student research projects) are currently in progress on the Ediacaran–Cambrian succession at Mount Dunfee. Please be respectful of ongoing research in this area, and please do not collect any specimens without first consulting the trip leaders.*

*Stop 3.1. Overview of Mount Dunfee*—The cars are parked near a red and white volcanic rock referred to by the locals as the "Elephant Rock."

This morning we will hike to the northeastern side of Mount Dunfee through the uppermost Reed Dolomite into the Esmeralda (middle) Member of the Deep Spring Formation (Fig. 9).

*Stop 3.2. Upper Reed Dolomite*—The Reed Dolomite is exposed in several fault blocks at Mount Dunfee. The tubular fossils *Wyattia* and *Nevadatubulus* have been reported from the upper member of the Reed Dolomite in the White-Inyo Mountains (Taylor 1966) and from the lowermost Deep Spring Formation at Mount Dunfee (Gevirtzman and Mount 1986, Signor et al. 1987). Although these taxa were initially thought to be Tommotian hyoliths, they were later synonymized with *Cloudina* by Grant (1990). Little detailed work has been done on the biomineralized taxa from the Reed Dolomite, largely because the degree of diagenetic alteration has eliminated all but general morphologic features. At Mount Dunfee, the uppermost beds of the Reed Dolomite contain ooids, pisoids, oncoids and biomineralizing tubular fossils in lenticular lag deposits (Gevirtzman and Mount 1986). The base of the Deep Spring Formation is placed above these beds at a prominent sequence boundary, which locally coincides with a paleokarst surface, and occurs in proximity to an irregular dolomization front.

This stop marks the entrance to the canyon in which we will walk up section. A composite measured section for the Mount Dunfee area was constructed from several fault blocks, two of which we will visit today (Figs. 7, 9). We will walk through sections E1425 and E1421 (Figs. 7, 9, 10). We will start in the southeast part of a small gully, at the base of E1425 in the upper Reed Dolomite. We will stop at a couple of outcrops of *Cloudina* around the corner from the entrance to the gully (Fig. 10C–D). At this locality and regionally, these fossils are preserved in lag deposits (Fig. 10C), with no convincing examples of *in situ* specimens.

*Discussion Questions*—

1. Are *Cloudina* the first metazoan reef builders? Are they living *in situ* in microbial reefs? Are they transported as detritus?
2. What is the FAD of *Cloudina* in the southwestern USA and globally?
3. Is the FAD of *Cloudina* something that could be used to subdivide the Ediacaran Period?

*Stop 3.3. Dunfee (lower) Member of the Deep Spring Formation: Ediacaran tubular body fossils and trace fossils*—We walked up the gully, still within the same fault block as we were at Stop 3.2, and have stopped at the base of the Deep Spring Formation and the lower of the two Ediacaran body fossil-producing intervals reported by Smith et al. (2016b) (Fig. 11).

The base of the Deep Spring Formation has been placed at a regionally distinct and easily recognizable sequence boundary that is defined by a sharp, karsted contact between a red, iron-rich, glauconitic, peloidal limestone and fine-grained siliciclastic facies (interbedded shale,

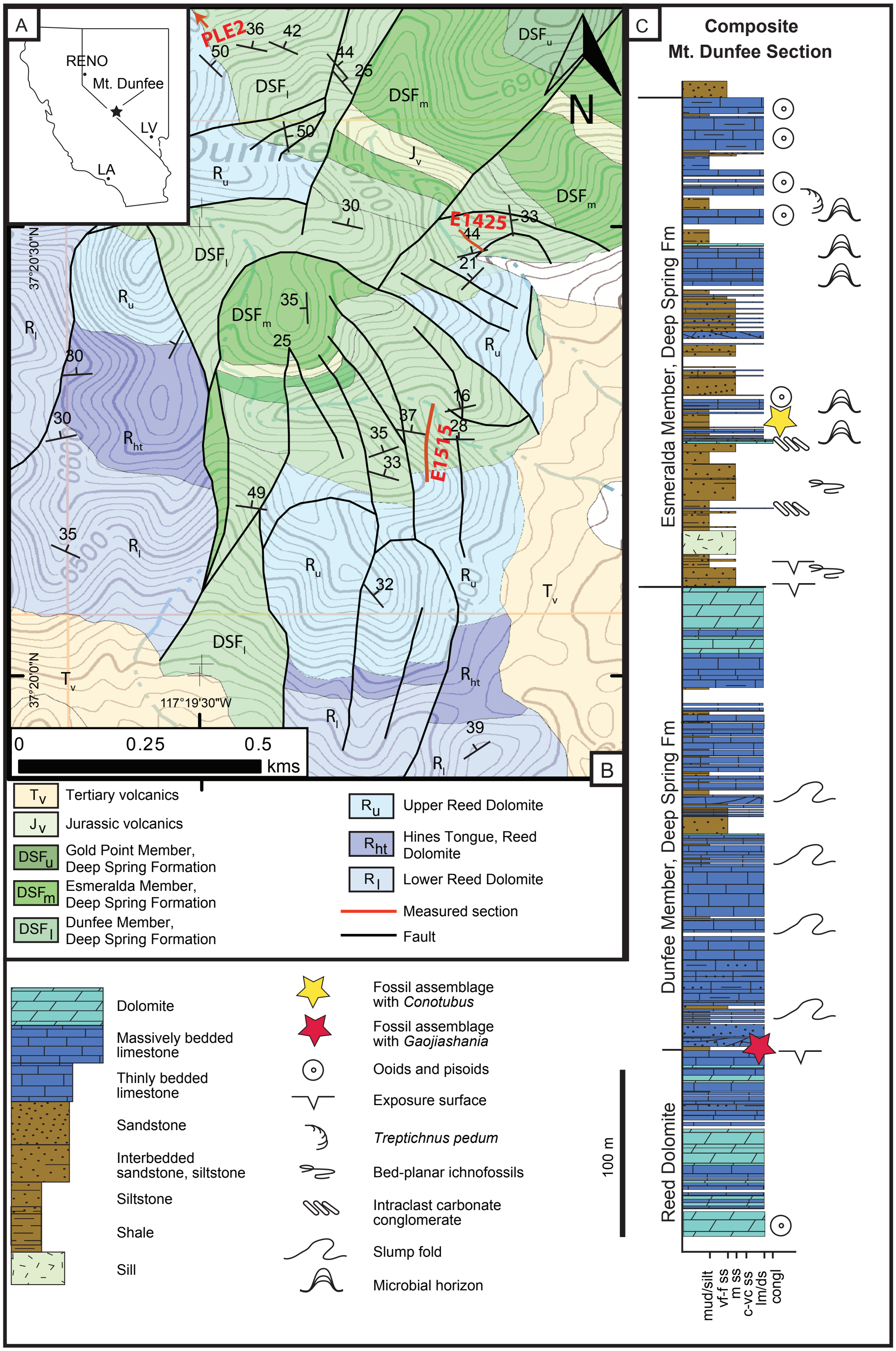



**Figure 7A–C.** Geology of Mount Dunfee. **A.** Locality map for Mount Dunfee, NV. **B.** Geologic map of Mount Dunfee. **C.** Composite measured section with biostratigraphy for the upper Reed Dolomite through Deep Spring Formation. Stratigraphic horizons of two Ediacaran body fossil assemblages are marked with stars. Modified from Smith et al. (2016b).


**Figure 8.** Satellite image of area surrounding Gold Point, NV showing stops for Day 3 of the field trip.

siltstone and sandstone) that contains mud chips, mud cracks, and syneresis cracks. Soft-bodied Ediacaran fossils that include *Gaojiashania*, *Wutubus* and smooth-walled pyritized tubular organisms have been found in this interval (Smith et al. 2016b) (Fig. 11). These fossils are preserved in a variety of taphonomic windows, including casts and molds, as two-dimensional pyrite pseudomorphs and associated with aluminosilicate minerals (Smith et al. 2016b) and have been collected from three different fault blocks at Mount Dunfee. Some of these fossils occur as poorly sorted fragments preserved on surfaces with mudstone chips and incipient muddy rip-up clasts, and thus we interpret them as having been locally reworked. More information on the *Gaojiashania* in this interval can be found in the section below.

*Discussion Questions*—

1. Are these tubular body fossils metazoans?
2. Is there evidence for vertical bioturbation prior to the FAD of *T. pedum*?

3.*3a:* Gaojiashania *at Mount Dunfee*—Upper Ediacaran successions around the world contain assemblages of 'tubular' macrofossils—fossils characterized by a tubular constructional morphology. Tubular fossils range from the remains of fully soft-bodied to lightly or even fully skeletonized forms and are represented by a wide range of taphonomic modes, from 'Ediacara-style' sandstone casts and molds to carbonaceous compressions to replacement by aluminosilicates, pyrite, phosphate or glauconite (e.g., Cai et al. 2010, Cai et al. 2012). Although the morphological differentiation and macroscopic size of tubular fossils indicates that they were likely multicellular eukaryotes, their precise affinities have remained largely elusive, with interpretations ranging from macrophytes to poriferan-, cnidarian- or bilaterian-grade animals. Given that the tubular body plan likely represents a convergent morphology, late Ediacaran tubular fossils may constitute a polyphyletic grouping (Tarhan et al. 2018).

The Deep Spring Formation at Mount Dunfee contains several assemblages of tubular fossils similar to those preserved in the upper Ediacaran Gaojiashan Biota of South China (e.g., Cai et al. 2010) (Fig. 11). The annulated tubular fossil *Gaojiashania* (Fig. 11A–B) is a common


**Figure 9.** Mount Dunfee field trip stops and walking path for the morning of Day 3 of the field trip.

component of assemblages in the Dunfee Member of the Deep Spring Formation, in which it occurs along multiple horizons, commonly with other tubular fossils such as *Wutubus*, as well as with bilaterian-produced trace fossils (see text below). *Gaojiashania* individuals in these assemblages occur on siltstone and sandstone bedding planes, are characterized by mm- to cm-scale diameters and appear to be preserved primarily as low-relief three-dimensional casts and molds, or two-dimensionally in aluminosilicate minerals. No holdfasts have yet been found associated with *Gaojiashania*, which may corroborate recent suggestions (Cai et al. 2013) that *Gaojiashania* had a procumbent life mode. Dunfee *Gaojiashania* moreover appear to have been mostly or entirely soft-bodied. Dunfee Member tubular and trace fossil assemblages occur in siltstone-dominated heterolithic siliciclastic packages interbedded with carbonate grainstone. Fossiliferous horizons contain mudstone intraclasts and desiccation cracks and immediately overlie a locally traceable paleo-karst surface and underlie HCS-bedded and locally scoured shoreface sandstones and shelfal carbonates, suggesting that Dunfee Member *Gaojiashania* lived in very shallow marine and likely peritidal settings (Tarhan et al. in preparation).

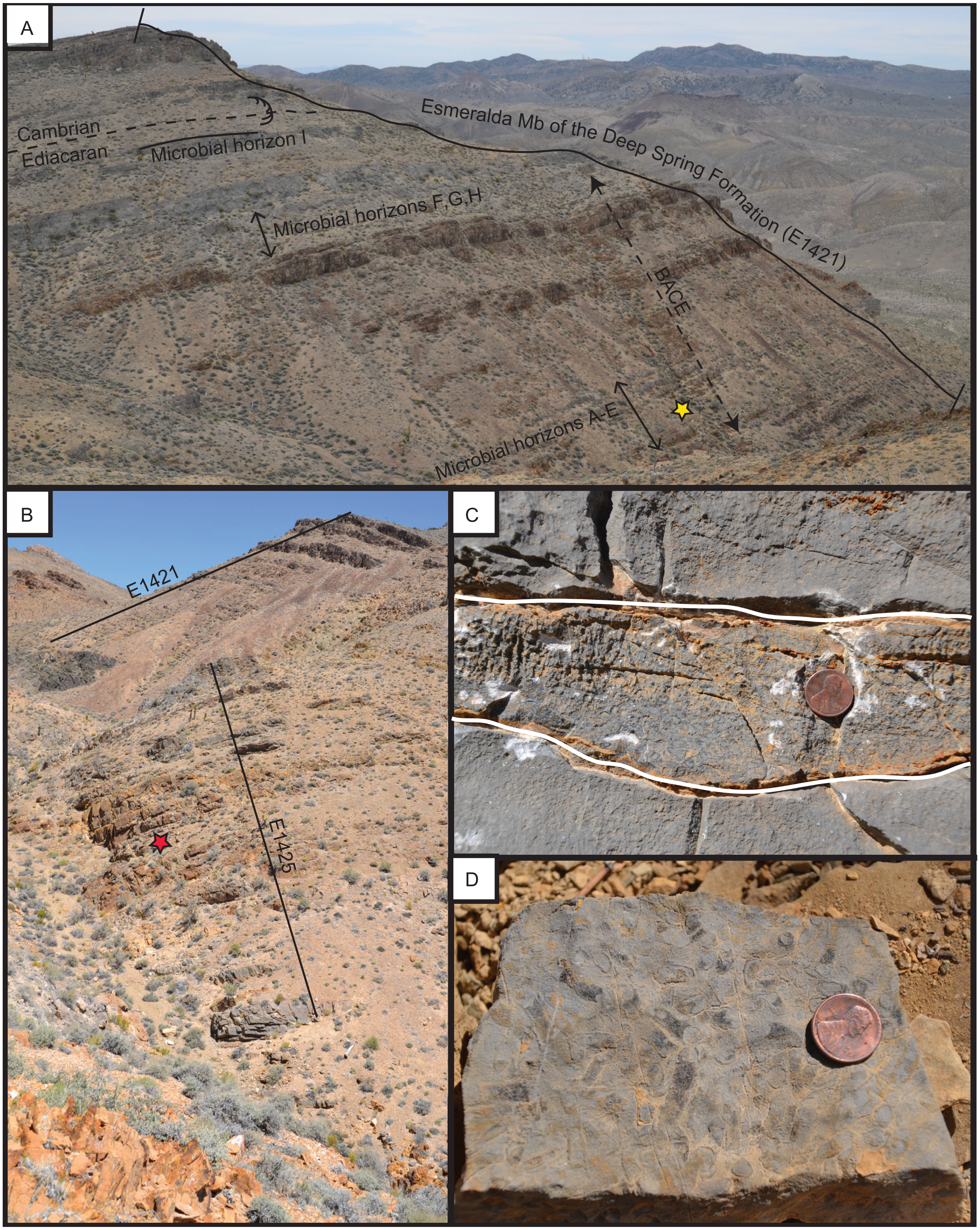



**Figure 10A–D.** Field photographs from Mount Dunfee. **A.** Photo of the Esmeralda (middle) Member of the Deep Spring Formation. The yellow star marks the *Conotubus* horizon. Lettered microbial horizons correspond to those in Rowland et al. (2008) and Figure 13. **B.** A fault block of the Dunfee (lower) Member of the Deep Spring Formation, with a separate fault block of the Esmeralda Member behind it. The red star marks the Dunfee Member fossil horizon. **C.** Lag deposit of *Cloudina* sclerites in the upper Reed Dolomite. **D.** Bedding surface covered by *Cloudina*. Modified from Smith and Nelson (in preparation).

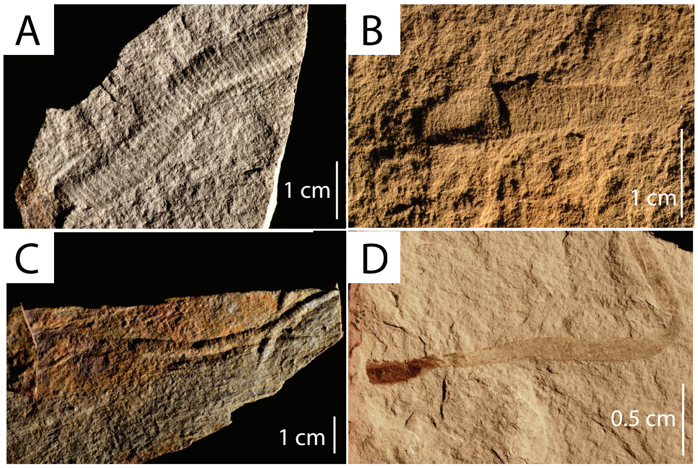
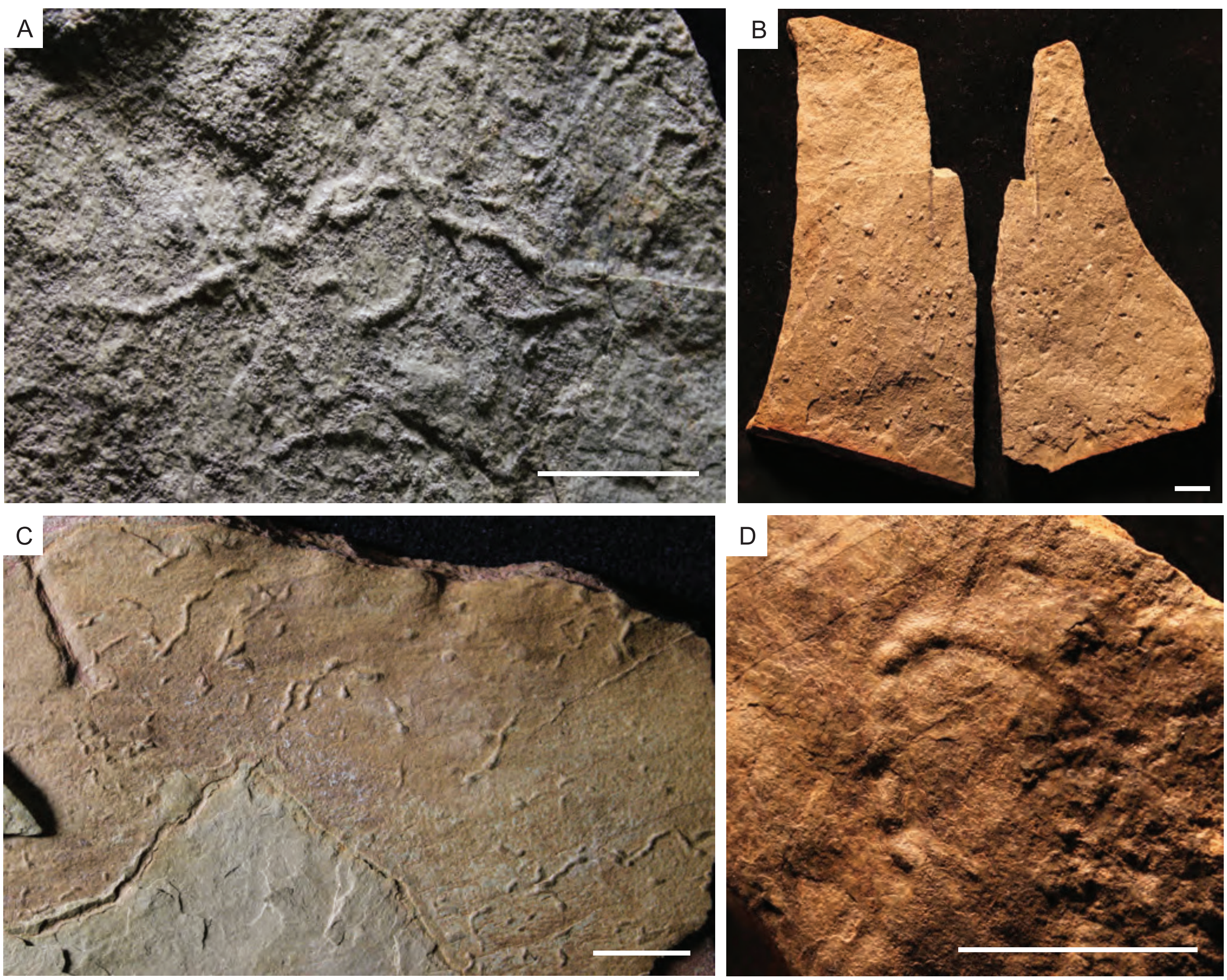



**Figure 11A–D.** Ediacaran body fossils from the lower Dunfee Member at Mount Dunfee. **A.** Cast of *Gaojiashania*. **B.** Cast and mold of *Gaojiashania*. **C.** *Wutubus*. **D.** Lightly pyritized smooth-walled tubular fossil. From Smith et al. (2016b).

*3.3b: Complex, sediment-penetrative trace fossils at Mount Dunfee*—The evolution of infaunal animals has dramatically shaped Earth's surface environments. Sediment-mixing (bioturbating) animals are, in modern marine environments, preeminent ecosystem engineers that physically and chemically rework seafloor sediments and thereby mediate nutrient cycling and the structure of benthic ecosystems on both local and ocean-wide scales.

The earliest sediment-displacement structures (i.e., trace fossils) created by bilaterian animals, simple curvilinear or meandering furrows, first appear in upper Ediacaran strata and record the advent of the infaunal lifestyle (e.g., Jensen et al. 2006). However, the majority of these structures are small, not significantly sediment-penetrative (of mm-scale diameters and depths) and although occasionally locally abundant, relatively uncommon by the standards of Phanerozoic trace fossil

**Figure 12A–D.** Trace fossils of the Dunfee Member of the Deep Spring Formation at Mount Dunfee. Scale bars=cm. Modified from Tarhan et al. (in review).



assemblages (e.g., Jensen et al. 2006). More morphologically complex and more deeply sediment-penetrative structures, recording behaviors characteristic of Cambrian infaunalization (e.g., Jensen et al. 2000, Jensen and Runnegar 2005, Chen et al. 2013, Macdonald et al. 2014, Meyer et al. 2014, Darroch et al. 2016, Buatois et al. 2018, Oji et al. 2018, Linnemann et al. 2019), as well as meiofaunal burrow networks (Parry et al. 2017), appear in uppermost Ediacaran strata. Documented examples of these Cambrian-style uppermost Ediacaran trace fossil assemblages nonetheless remain relatively uncommon; not until the early Phanerozoic do assemblages of morphologically and ethologically complex and large-scale trace fossils become relatively ubiquitous.

The Ediacaran succession at Mount Dunfee contains a variety of trace fossils, including simple meandering furrows (cf. *Helminthoidichnites*, *Helminthopsis*) and shallowly emplaced infilled burrows (cf. *Planolites*) which are preserved along bedding planes of the heterolithic Dunfee (lower) and Esmeralda (middle) members of the Deep Spring Formation (Gevirtzman and Mount 1986). These trace fossils are particularly common throughout the siliciclastic-dominated lowest interval of the Dunfee Member, as well as in the overlying Esmeralda Member. These structures are of millimetric diameters and depths and are not fabric-disruptive; the beds along which they occur are commonly well-laminated (Tarhan et al. in review). Additionally, Dunfee Member trace fossil assemblages include structures of more complex morphology, including regularly sinuous (cf. *Cochlichnus*), spiraling (cf. *Helicolithus*) and treptichnid-like forms similar to trace fossils observed in upper Ediacaran strata of Finnmark (Högström et al. 2013, McIlroy and Brasier 2017), as well as bed-penetrative dimpled plugs (cf. *Bergaueria*) (Tarhan et al. in review) (Fig. 12). Although these traces, like the morphologically simpler forms with which they co-occur, are of millimetric scale, they are remarkable for their relative complexity and the frequency with which they occur along siliciclastic bedding planes of the lower Dunfee Member (Tarhan et al. in review). This is particularly striking given that these trace fossil assemblages occur over 500 m below the local Ediacaran–Cambrian boundary as previously defined at the first occurrence of *T. pedum* (Corsetti and Hagadorn 2003) and nearly 400 m below the local expression of the nadir of the basal Cambrian negative carbon isotope excursion (BACE) (Smith et al. 2016b). This assemblage may therefore provide one of the oldest documented records of simple sediment-penetrative infaunalization (Tarhan et al. in review). Moreover, these Dunfee Member trace fossil assemblages occur along the same horizons as classic upper Ediacaran tubular fossils such as *Gaojiashania* and *Wutubus*, suggesting that ecologically sophisticated, although low-diversity, communities of epifauna and infauna may have been common in these shallow marine settings.

*Stop 3.4. Esmeralda (middle) Member of the Deep Spring Formation: Ediacaran body fossils and stromatolites*—We just walked to the base of the ridge in Figure 10A, and have stopped at the first carbonate bed. The Esmeralda Member Deep Spring stromatolites are a frequent field trip stop (i.e., Rowland et al. 2008) and have been known as one of the classic 'Rowland's Reefs' stops for over a decade (Fig. 13). This section has been identified as one of the best places to examine morphological reef diversity in latest Ediacaran reefs. More recently, the late Ediacaran body fossil *Conotubus* (Fig. 14) and the algal fossil *Elainabella* have been discovered in siltstone interbedded with these lowermost reefs and within the onset of the BACE, establishing these fossils as some of the youngest known Ediacaran fossils globally (Rowland and Rodriguez 2014, Smith et al. 2016b).

*3.4a: Esmeralda Member microbialites and depositional environment at Mount Dunfee*—Overlying the siliciclastic-dominated strata in the lower part of the Esmeralda Member are strata containing oolites and microbialites (Figs. 7, 10A, 13). These carbonate-dominated strata can be divided into two intervals, lower and upper, that complete shoaling-upward cycles within the section, and have been interpreted as representing peritidal reefs and shoals. The first cycle is capped by the carbonate interval containing microbial horizons A–E (Figs. 7, 10A, 13). The base of the second interval is dominated by mixed carbonate-siliciclastic shallow subtidal strata which are capped by microbial horizons F–I and ooilites (Figs. 7, 10A, 13). Within some of the microbial intervals—particularly horizons C–E—meter-scale cyclicity is recorded.

In general, the depositional environment of the Esmeralda Member at Mount Dunfee is thought to be a narrow, discontinuous, episodically emergent reef and shoal complex on a ramp (Oliver 1990). This section represents a geographically restricted locus of carbonate sedimentation and microbialite reef construction at the very end of the Ediacaran Period.

*3.4b:* Conotubus *at Mount Dunfee*—Large numbers of conical and adapically tapering tubular fossils have been recovered from the Esmeralda Member of the Deep Spring Formation (Fig. 14; Smith et al. 2016b). These structures consist of nested cylindrical to funnel-shaped

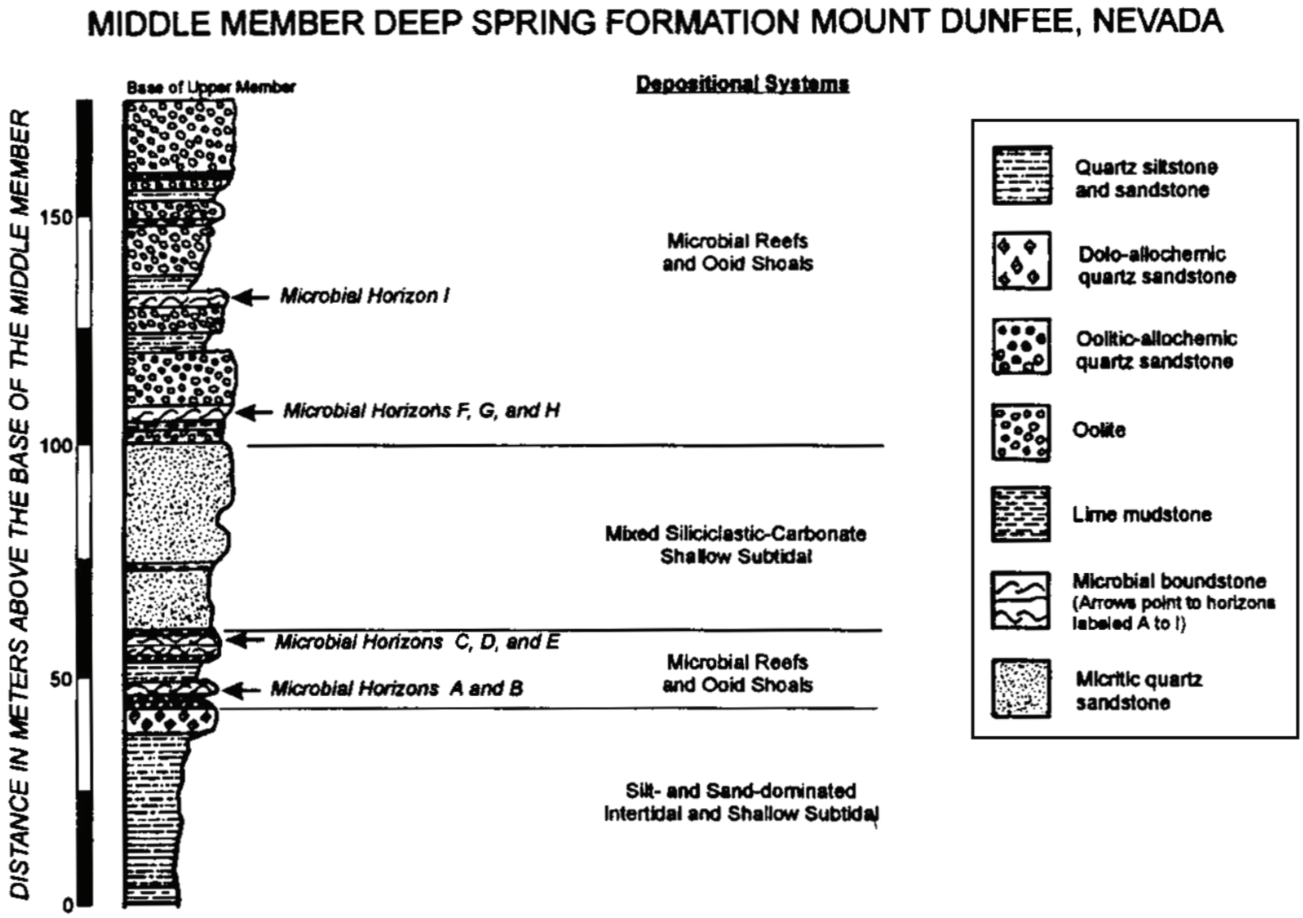



**Figure 13.** Stratigraphic column of the Esmeralda (middle) Member of the Deep Spring Formation at Mount Dunfee, depicting the lettered microbial horizons, and interpreted depositional systems, shown in Figure 10A. From Oliver (1990).

structures and are preserved three-dimensionally as iron oxides, interpreted as the oxidized remnants of pyrite. On the basis of morphological and taphonomic similarities, these fossils have been referred to *Conotubus*, previously known solely from the Gaojiashan Member of the Dengying Formation of South China (e.g., Cai et al. 2011). More recently, pyritized *Conotubus* have also been discovered in the lower member of the Wood Canyon Formation in the Montgomery Mountains of the Death Valley region (Smith et al. 2017). The taxonomy and taphonomy of the Mount Dunfee *Conotubus* are currently being studied in more detail by paleontologists at University of Nevada, Las Vegas and University of Missouri.

*Conotubus* has been reconstructed as an erect epibenthic organism with a suspension-feeding life mode and the ability to rejuvenate and self-correct its growth direction following storm-mediated felling and burial (Cai et al. 2010, Cai et al. 2011). Due to this apparent morphological plasticity and evidence for non-brittle deformation, and the lack of evidence for strongly biomineralizing wall structure, *Conotubus* has been inferred to have been soft-bodied or only lightly biomineralizing (e.g., Cai et al. 2011, Wood et al. 2017). Some workers have highlighted the morphological similarity between *Conotubus* and *Cloudina* and suggested that these taxa may be related, with the biomineralized *Cloudina* potentially recording acquisition of the potential for environmentally-mediated skeletonization by a *Conotubus*-like precursor. However, at Mount Dunfee, unlike in the Dengying Formation of South China, the local LAD of *Cloudina* stratigraphically underlies the local FAD of *Conotubus*.

### Day 3 (6/21) afternoon: Archaeocyath reefs at Stewart's Mill

*We will spend midday and have lunch at the nearby pseudo-ghost town of Gold Point. There will be the opportunity to purchase cold beverages (cash only).*

*There will be an optional visit in the afternoon to the*

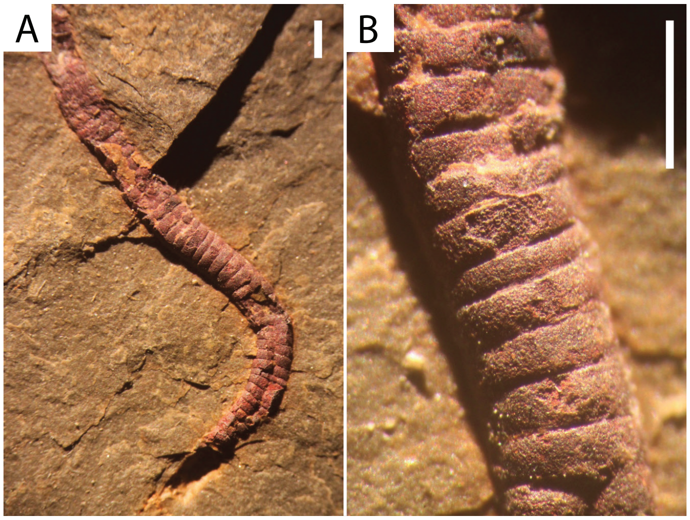



**Figure 14A, B.** Tubular body fossils of the Esmeralda Member of the Deep Spring Formation at Mount Dunfee. **A, B.** Two images of a single pyritized *Conotubus* specimen, now preserved as iron oxides. Scale bar=0.1 cm. From Smith et al. (2016b).

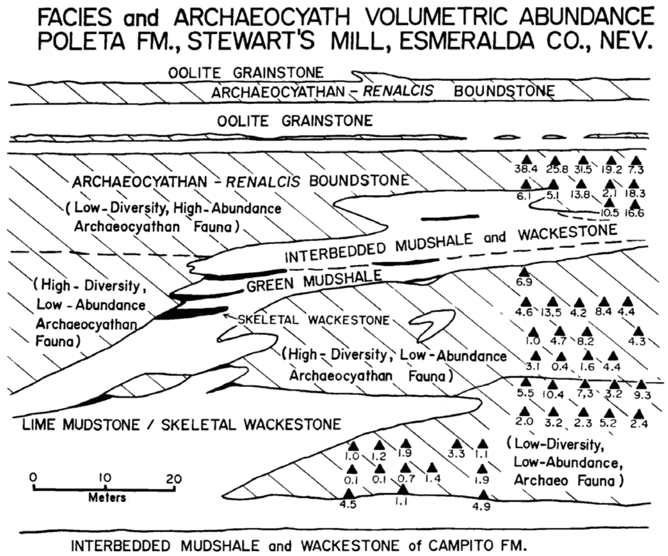



**Figure 15.** Facies relationships within the reef-lagoon interval at Stewart's Mill (Stop 3.5). Numbers below the black triangles indicate the volumetric abundance of archaeocyaths. From Rowland and Gangloff (1988).

*archaeocyathan reefs of the lower Poleta Formation, at Stewart's Mill.*

*Stop 3.5. Archaeocyathan-Renalcis reefs of the Poleta Formation at Stewart's Mill—*

*Note: This exposure is one of the best examples in the world of early Cambrian archaeocyathan reefs. Researchers from all over the world come here to study reef complexes, and many faculty bring their students here. If you want to collect specimens at this stop, please collect samples only from float.*

A few kilometers northwest of Gold Point, NV, an outcrop, known as Stewart's Mill (and, due to extensive work on this succession by Steve Rowland and his colleagues, also widely known as one of "Rowland's Reefs") contains beautiful exposures of an archaeocyath reef complex in the midst of oolite shoals (Fig. 15). (Rowland 1984, Rowland and Gangloff 1988, Rowland and Hicks 2004, Rowland et al. 2008). The below text and figures were modified from Rowland et al. (2008).

The lower third of this outcrop consists of relatively poorly exposed green shales, siltstones and minor interbedded sandstones and carbonates of the Montenegro Member of the Campito Formation. The middle portion of the outcrop, which comprises the lower member of the Poleta Formation, consists of a 70 m-thick package of thrombolitic and archaeocyathan-*Renalcis* boundstone. This boundstone-dominated interval is overlain by a cliff-forming unit of oolite, biostromes and packstone (Fig. 15).

This reef succession consists of the following facies, in ascending order (though complex interfingering is common; Fig. 15): (1) bioclastic carbonate mudstone, (2) thrombolites with sparse archaeocyaths, (3) *Renalcis*-dominated boundstone, (4) green mudstone with lenses of skeletal wackestone, (5) archaeocyath-dominated boundstone, and (6) oolite grainstone, which are interpreted to record reef and lagoonal settings (Rowland et al. 2008). The green mudstones with skeletal wackestones of Facies 4 have been reconstructed by Rowland et al. (2008) as a bypass channel through which siliciclastic sediment was transported across the margin of the carbonate shelf. The fossiliferous facies in this succession appear to be ecologically zoned (Rowland et al. 2008).

The lower portion of this package is dominated by thrombolites, with archaeocyaths (largely unbranched varieties) comprising <3% of total rock volume, whereas the uppermost portion of this package is dominated by a high abundance of branching archaeocyaths, which comprise up to 38% of total rock volume (Rowland et al. 2008). This zonation scheme is summarized in Figure 16. Over this interval, there are conspicuous changes in (1) the relative abundance of archaeocyaths, which increases upsection; (2) the morphology of archaeocyaths, with non-branching forms dominant throughout most of this interval and branching forms becoming dominant within the uppermost 30 m; and (3) the relative abundance of reef cavities; decimeter- and meter-scale cavities (preserved as dolomite-rich, orange patches) are absent in the lower third, but are conspicuous in the upper

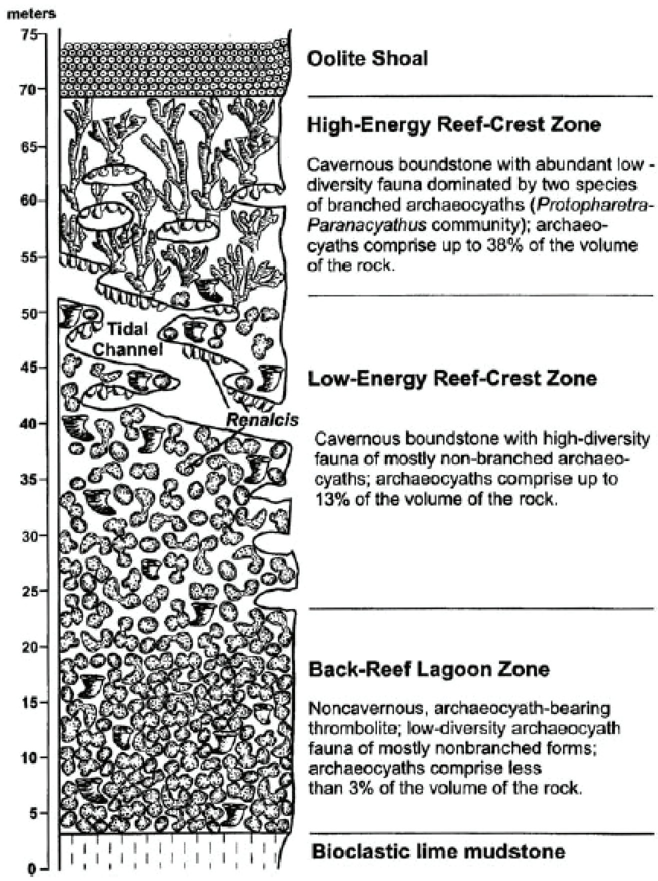



**Figure 16.** Vertical ecological zonation in the reef-lagoon facies complex of the lower member of the Poleta Formation at Stewart's Mill (Stop 3.5). From Rowland and Hicks (2004).

two-thirds of this interval. Similar patterns of zonation have been described in fossil reefs of a variety of ages, including Ordovician, Silurian, Devonian, and Cretaceous examples (Walker and Alberstadt 1975). Walker and Alberstadt (1975) have interpreted such zonation as a record of ecological succession. Alternatively, Rowland and Gangloff (1988) have interpreted this zonation as the product of facies migration during an interval of marine transgression; Rowland and Shapiro (2002) subsequently divided this package into three environmental zones: (1) a lower back-reef lagoon zone, (2) a middle low-energy reef-crest zone, and (3) a high-energy reef-crest zone, which is distinguished from the underlying lagoonal facies by the presence of infilled cavities, and from the underlying low-energy reef-crest facies by the presence of abundant branching archaeocyaths (Fig. 16).

*Return to the hotels in Beatty. Dinner will be at the Sourdough Saloon, a (0.6 km) walk from our hotels. Participants are responsible for paying for their own dinners.*

## Day 4 (6/22): Cryogenian–Ediacaran stratigraphy of the Death Valley region

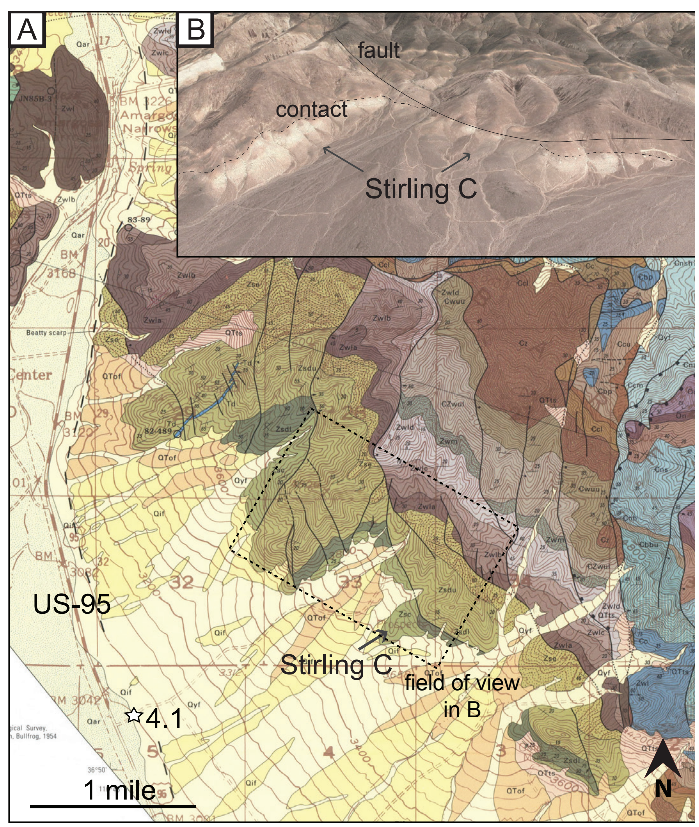
*Participants responsible for their own breakfasts; check out and depart hotels by 8:00 am.*

*We will lay out lunch supplies in the morning. Please pack a lunch prior to departing Beatty. Participants should eat their packed lunches at their discretion; we will be making several brief stops in the morning and early afternoon, but we will not be making a formal lunch stop. We recommend eating your lunch in the car.*

*We will be visiting terrain that is close to sea-level and thus experience even hotter temperatures today. All stops today will be brief and along the road. Nonetheless, please take care to stay hydrated and be mindful of sun exposure.*

*Stop 4.1: Quick stop on I-95 to look at strata exposed on Bare Mountain, just south of Beatty*—This is a quick stop on I-95 to discuss the transition from the White-Inyo region stratigraphy to Death Valley region stratigraphy. Just north of Beatty, NV, marks the transition from the more paleo-distal stratigraphic deposits of Esmeralda

**Figure 17A–B.** Ediacaran stratigraphy of Bare Mountain (Stop 4.1). **A.** Geologic map of Bare Mountain (Monsen et al. 1992). Dashed box marks area of satellite image shown in panel B. **B.** Satellite image highlighting the carbonate-rich C member of the Stirling Quartzite.



County and the White-Inyo Mountains to more proximal correlative units of the Death Valley region. The succession at Bare Mountain (to the east of the highway) contains one of the few sections of the Stirling Quartzite that is carbonate-dominated in the middle–upper part of the unit, and contains facies more similar to those of the Reed Dolomite than do most other sections of the Stirling Quartzite (Fig. 17). At this locality, the metamorphic grade of the Stirling is, however, higher than at most other sites in the Death Valley region, leaving the dolomite sucrosic and recrystallized.

If you look to the west, you will see the Funeral Mountains in the distance, which are part of the Death Valley region. These sections also contain facies transitional between the Ediacaran–Cambrian stratigraphies of the White-Inyo Mountains/Esmeralda County and the Death Valley region (Stewart 1970). *Cloudina* has been reported from a Funeral Mountains section of Member D of the Stirling Quartzite (Langille 1974).

*Stop 4.2: Pliocene–Pleistocene Lake Tecopa—We will have a 30-minute rest stop and brief discussion in the town of Shoshone. There are public toilets and cold beverages or food items can be purchased from the general store.*

Lake Tecopa is a Pliocene-Pleistocene basin-fill sequence. Mudstone, sandstone, and conglomerate interbedded with tuff accumulated between 5 Ma and 186 Ka. Some of the intercalated tuff beds have been chemically correlated with units recording well-known eruptions in the western United States, such as the Bishop Tuff (759 Ka) and the Lava Creek Tuff (639 Ka) (Izett 1981, Sarna-Wojcicki et al. 2000). The sedimentary sequence has been subsequently cut by late Pleistocene incision of the Amargosa River (Morrison 1999).

The Tecopa Basin has been interpreted as a saline lake that was surrounded by springs and fluvial inlets (Gibert et al. 2011). The lake deposits record fluctuations in water level and salinity, which resulted in several thick gypsum horizons, which were subsequently mined in the early twentieth century. Some of these mining camps were located right around Shoshone in an area known as Dublin Gulch. Due to a scarcity of housing and building material, some of the miners carved cave dwellings into the Tecopa Lake beds and tuff horizons. These dwellings can be seen on the right (west) as we drive out of

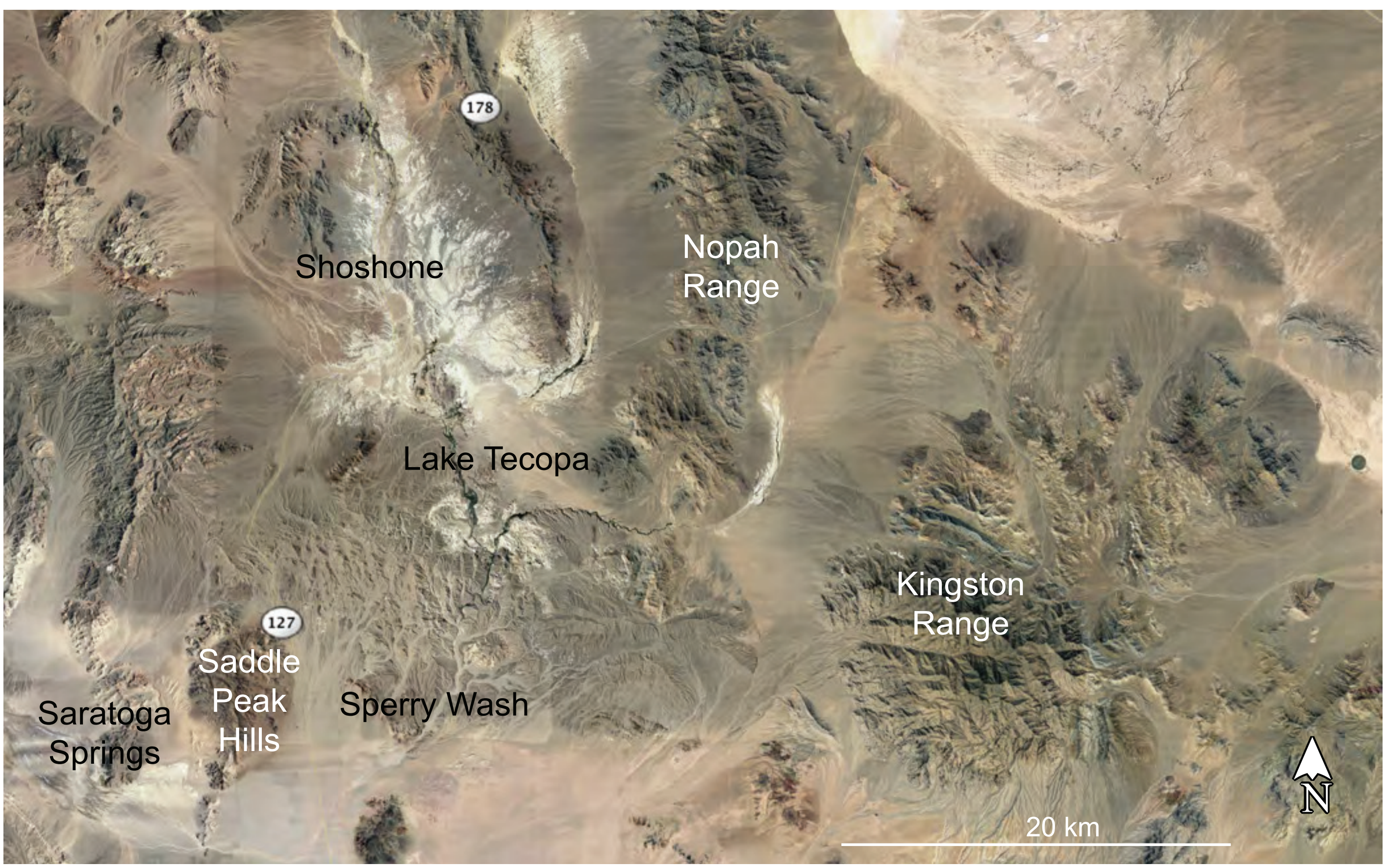



**Figure 18:** Satellite image of southeast Death Valley showing many of the stops of Day 4 of the field trip.

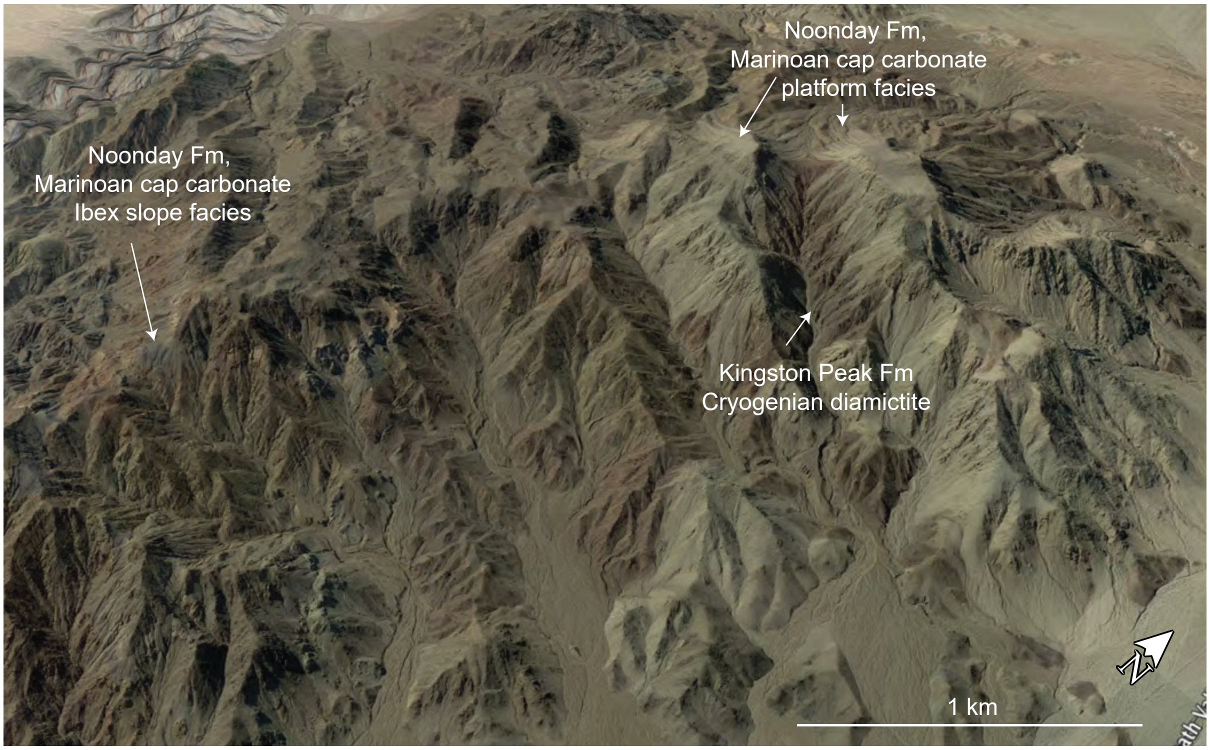



**Figure 19:** Satellite image of Saddle Peak Hills showing the platformal (Nopah) facies and basinal (Ibex) facies of the Noonday Formation.

Shoshone.

*Stop 4.3: Cryogenian strata in the Saddle Peak Hills and along Sperry Wash Road—*

*4.3a: Saddle Peak Hills*—The Saddle Peak Hills are about 40 km south of Shoshone, just west of Route 127 (Fig. 18, 19). The hills are named for the prominent saddle between the two tallest peaks that are capped by the Noonday Formation. These Cryogenian exposures are not as frequently visited as those in other parts of Death Valley (i.e., the Nopah Range and Kingston Range), but they offer a glimpse into different facies of the Tonian–Ediacaran succession of Death Valley. Specifically, this is one of the few places to see, in a single transect, the transition from reef to slope facies of the Noonday Formation cap carbonate.

*Discussion Questions—*

1. How continuous is the Cryogenian–Ediacaran record in Death Valley?
2. Where are there major unconformities?

4.3b: *Sperry Wash*—Sperry Wash is one of the best places in the Death Valley region to see evidence of ice-rafted debris of Cryogenian glacial deposits. At this stop (Fig. 18), we will see the Cryogenian Kingston Peak Formation and the overlying Noonday Formation, the lowest member of which has been correlated to the Marinoan cap carbonate (Prave 1999, Petterson et al. 2011), dated globally at 635 Ma (Hoffman et al. 2004, Condon et al. 2005, Calver et al. 2013). At most localities of the Kingston Peak Formation, dropstones and other sedimentary indicators of glaciation are rare.

*Discussion Questions—*

1. Are the glacial deposits in the Kingston Peak Formation Sturtian, Marinoan or both?
2. Do the outsized clasts in the Kingston Peak Formation necessitate a glacial origin?

*Return to University of California, Riverside. We will make a brief stop* en route *at Baker, California, to refuel the vehicles and use the restrooms.*

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