PALEOECOLOGY OF UPPER CAMBRIAN CARBONATE

UNITS, HOUSE RANGE AND SOUTH LAKESIDE

MOUNTAINS, UTAH

by

Patricia K. Garcia

A thesis submitted to the faculty of
The University of Utah
in partial fulfillment of the requirements for the degree of

Master of Science

in

Geology

Department of Geology and Geophysics

The University of Utah

December 2015

The University of Utah Graduate School

STATEMENT OF THESIS APPROVAL

The thesis of   **Patricia K. Garcia**

has been approved by the following supervisory committee members:

Allan A. Ekdale	, Chair	**10/02/2015**
		Date Approved
Marjorie Chan	, Member	**10/02/2015**
		Date Approved
Randall B. Irmis	, Member	**10/02/2015**
		Date Approved

and by   **John M. Bartley**   , Chair/Dean of

the Department/College/School of   **Geology and Geophysics**

and by David B. Kieda, Dean of The Graduate School.

ABSTRACT

Upper Cambrian strata in the Great Basin of western Utah are dominated by marine carbonate facies. The Hellnmaria Member of the Notch Peak Formation in the southern House Range and the Upper Carbonate Member of the St. Charles Formation in the northern South Lakeside Mountains are correlative Upper Cambrian strata. At both localities, the correlative carbonate units preserve cyanobacterical stromatolite facies containing burrows.

The Hellnmaria Member and the Upper Carbonate Member both contain smooth walled cylindrical burrows inside and surrounding the stromatolites. The burrows exhibit a branching morphology, and they have a coarser grain fill that is different in texture from surrounding sediments. The branching geometry of the burrow tunnels is irregular and constricted. These are assigned to the trace fossil *Thalassinoides paradoxicus.*

Thalassinoides paradoxicus is a deep infaunal deposit feeding trace fossil (fodinichnion) constructed by burrowing organisms of unknown identity. The probable burrowers mined shallow marine muds for food as well as seeking protection in the complex web of burrows. The *Thalassinoides* of this study are far too old to have been created by decapod crustaceans, which are the primary creators of such trace fossils in the Mesozoic and Cenozoic, so the identity of the Cambrian trace makers remains unknown.

Many of the *Thalassinoides* burrows have been replaced with chert, which appears to be early diagenetic in origin when the burrows were still porous enough for the silica to find favorable nucleation sites on the burrow surface. Many of the burrows are over-silicified due to growth of the microcrystalline chert nodules extending beyond the boundaries of the original trace fossil.

The Hellnmaria and Upper Carbonate members represent a shallow, tropical, quiet water paleoenvironment on an open marine platform. This interpretation is due to the presence of stromatolites and *Thalassinoides* burrows, both of which flourished in this paleoenvironment.

This study demonstrates the importance of trace fossils in the interpretation of paleoecology, paleoclimate, and paleoenvironment of the ancient carbonate strata during the Late Cambrian in the Great Basin. This is particularly useful, because much of the strata of this age in this region lack a substantial body fossil record.

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CHAPTER 1

INTRODUCTION

1.1 Purpose and Significance

Laminated microbial stromatolite facies were widespread during the Cambrian-Ordovician transition in the eastern Great Basin of western Utah. The stromatolites contain a heterogeneous sedimentary fabric exemplified by structures interpreted variously as burrows (trace fossils) or inorganic diagenetic structures. A full picture of the benthic community in ancient stromatolite habitats is poorly understood; therefore, fundamental questions need to be answered about Cambrian stromatolite facies before meaningful paleoecology research can be accomplished.

Widespread ecosystems during the Cambrian were based on warm carbonate paleoenvironments. The paleoecology of Upper Cambrian paleocommunities is known, in part, from such body fossils as primitive mollusks, trilobites, and conodonts. Burrow-like structures within and between microbial stromatolites of the Late Cambrian found in the southern House Range and the South Lakeside Mountains of Western Utah offer important, new insights into the Cambrian paleoecology and paleoenvironments of the microbial stromatolite facies have been revealed. This study presents comprehensive evidence to show that these structures, located within and between individual stromatolites, are burrows (*Thalassinoides paradoxicus*).

1.2 Research Questions

A central question in this study is to assess whether or not certain secondary structures that can be observed between and inside the Upper Cambrian stromatolites are burrows.

This study also investigates the paleoenvironmental setting during the Late Cambrian in what is now the Great Basin of Utah. This study addresses the paleoclimatic implications of the ichnology and paleoecology in this area during the Late Cambrian.

1.3 Hypotheses

The principal hypotheses are: 1) structures inside and between stromatolites are *Thalassinoides* burrows made by infaunal deposit feeders; 2) the stromatolite facies represents a tropical paleoenvironment with shallow, quiet, and perhaps highly alkaline water; and 3) trace fossils are important information in the interpretation of paleoecology, paleoclimate, and depositional environments of the ancient carbonate strata during the Late Cambrian in the Great Basin, because most of the strata of this age and region lack a substantial body fossil record.

1.4 Focus of Study

This study focuses on Late Cambrian paleoecology and ichnology in the Great Basin of Utah. Identifying and defining the burrow-like structures is a crucial part of a comprehensive analysis of the Great Basin. Naming the burrows in this field area is important for clarification and comparison to other known occurrences of this particular type of burrow system.

Behavioral convergence of trace-making animals must be addressed in this study if the burrow-like structures are *Thalassinoides*. This type of burrow is almost exclusively attributed to decapod shrimp, but the burrows examined for this study are approximately 120 million years older than the first occurrences of decapod body fossils (Feldmann and Schweitzer, 2010).

Subsequent chapters focus on the paleoecology, paleoenvironment, and paleoclimate of the stromatolitic facies that make up the Hellnmaria Member of the Notch Peak Formation and Upper Carbonate Member of the St. Charles Formation in the Southern Lakeside Mountains and the southern House Range. The ichnology of the area provides pertinent evidence about the life and environment of this area during the Late Cambrian.

1.5 Cambrian Exposures

Cambrian and Ordovician strata in the two research areas are located within the Basin and Range Geologic Province in western Utah, also known as the Great Basin. Both study areas contain the carbonate rock units with stromatolite facies. The exposures of the Upper Cambrian dolomite and limestone units crop out for long distances in the research area, with dip angles as low as 10° in places (Hintze and Kowalis, 2009). The South Lakeside Mountains are located along the southwestern side of the Great Salt Lake in Toole County, and the southern House Range is located southwest of Sevier Lake in Millard County.

The Cambrian paleogeography in the Great Basin has been divided into three zones: an inner detrital belt, a middle carbonate belt, and an outer detrital belt (Figure 1.1;

Palmer, 1971). The South Lakeside Mountains and the southern House Range areas both contain well-exposed Upper Cambrian strata representing the middle carbonate belt (Figure 1.2).

1.6 Depositional Patterns in the Cambrian

The Cambrian Period began approximately 541 Ma and lasted about 56 million years (Gradstein et al., 2012). The Cambrian Period was characterized by global transgressive intervals when the western ocean expanded eastward onto the shelf and across the continent (Stewart, 1992). The deposition during the Middle to Late Cambrian was influenced by major sea level fluctuations in the Cordillera (Burchfiel et al., 1992). By the Late Cambrian, the shoreline extended as far to the east as the Black Hills area of South Dakota (Figure 1.3). The western part of the sea gradually deepened during the Cambrian, and the sedimentary deposits grew thicker westward into Utah (Burchfiel et al., 1992; Stokes, 1986). In the southern Great Basin of Utah, the Lower Cambrian quartzite facies changed to siltstone, and finally to almost entirely shallow-marine carbonates in the Upper Cambrian units (Stewart, 1992).

The Cambrian was characterized by some of the highest sea levels of the Phanerozoic Eon. Widespread carbonate platforms developed during this time, some of them hundreds of miles wide (Cook, 1992; Poole, 1992). Extensive platforms such as those existing during the Cambrian have no modern analogue, a difficulty that hinders attempts to interpret the Cambrian paleoenvironment. The Upper Cambrian platform deposits are predominately shallow-subtidal carbonate rocks in units up to 2,000 m thick (Poole, 1992). During this period, atmospheric carbon dioxide levels reached as high as

6000-7000 ppmv according to GEOCARB III models (Berner, 2009). These high levels of atmospheric carbon dioxide may have created strong greenhouse forcing, producing a warm climate that precluded continental ice and may account for high eustatic sea level at this time.

1.7 Stromatolites

Large numbers of microbial stromatolites formed in this area during the Late Cambrian (Figure 1.4). Stromatolites are biostratification structures created by the interaction of cyanobacteria and sediment in water. Because grazing by metazoan herbivores tends to destroy stromatolites, the Late Cambrian may have been an interval of reduced metazoan activity, possibly due to a rapid change in sea level and/or climate conditions, or a change in pH of the seawater.

Modern stromatolites are known to form only in extreme conditions where other organisms, such as most algal grazers, cannot survive. Present-day microbial mats are heavily grazed by herbivores, such as archeogastropods and other invertebrates. Large domal microbial structures resembling those that were widespread in the Early Paleozoic are restricted to such locations as Shark Bay, Australia, and Lee Stocking Island, Bahamas, where invertebrate diversity is reduced by seasonal hypersalinity and mobile sediment in the current swept channels (Feldman and McKenzie, 1998). The possibility that consumption by metazoans has reduced the extent of microbial carbonates and progressively relegated them to ecological refuges is shown by their broad pattern of distribution (Awramik, 1971).

1.8 Regional Geology

The two research areas are located in the Basin and Range geologic province in western Utah. The South Lakeside Mountains and the southern House Range areas were chosen for study, because they contain what appears to be a heavily burrowed stromatolite facies in correlative Upper Cambrian strata (Figure 1.5) (Figure 1.6).

The study areas both reveal Lower Cambrian strata resting unconformably on Neoproterozoic siliclastics and structurally complex basement rocks (Stewart, 1992; Hintze and Kowallis 2009). The Upper Cambrian Notch Peak Formation in the House Range is up to 422 m thick (Miller et al., 2011) and contains the Hellnmaria Member, which exposes a stromatolitic facies exposed in several areas, including the principal collection site on the Wah Wah Arch near Steamboat Pass in the southernmost part of the House Range. The lower half of the Hellnmaria Member at Steamboat Pass is massive limestone and the upper section is dolomite (Miller et al., 2011). The stromatolitic facies at Steamboat Pass, where samples were collected, has been determined to be dolomite by QEMSCAN and thin sections. This facies includes ooids, oncoids, and columnar stromatolites.

The correlative rocks of the South Lakeside Mountains are similar in character to those in the southern House Range. The Upper Carbonate Member is a stromatolitic facies that has been dolomitized. Because of the continuous rises of sea level through the Cambrian, the Upper Carbonate Member overlies a transgressive sequence of basal sandstones. The Upper Cambrian dolomite has overlying units of laminated dolomite that contains no stromatolites or burrows and is a tan color instead of the blue-gray color of the Upper Carbonate Member. The overlying Ordovician rock units of the Garden City

Formation represent a drop in sea level.

1.9 Regional Geologic History

The Precambrian supercontinent Pannotia broke up during the Neoproterozoic and produced a newly rifted margin beginning in central and eastern Nevada and forming the edge of Laurentia (Sears and Price, 2003). A tropical shallow-marine carbonate platform formed after the rifting, and along the rifted margin, there was rapid subsidence, resulting in accommodation space for a thick sequence of continuous marine deposits (Stewart, 1992). These marine deposits resulted in a 3.87 km thick package of Cambrian strata (Hintze and Kowalis, 2009; Miller et al., 2011). The ocean gradually advanced onto the continent and by the middle Late Cambrian had reached maximum water depth (Ziegler et al., 1979).

In the Late Cambrian, a north-trending carbonate belt extended along the continental margin. Most landmasses during this time were situated at low latitudes, while the high latitudes were mostly covered by ocean (Sears and Price, 2003). There are deposits representing the continental margin, continental slope, and continental rise sediments along eastern Utah. Deposition continued into deeper water to the west in Nevada, resulting in carbonate and siliclastic deposits (Taylor and Cook, 1976). Farther to the southwest is a continuation of the same carbonate bank from western Utah and southeastern Nevada (Kepper, 1981).

The Cambrian strata that were studied for this project experienced eastward thrusting during the Cretaceous Sevier Orogeny. Then in the Cenozoic, Basin and Range extensional faulting began producing the horst and graben pattern observed in this area

and the majority of western Utah today. In some parts of this area, Quaternary basalt flows covered the older deposits (Miller et al., 2012).

1.10 The Southern House Range

During the Late Cambrian and Early Ordovician, the House Range was tectonically active, and the southern area of the range was bounded by a fault. The loading of Cambrian and Ordovician sediments and continued subsidence facilitated the fault movements, which aided in the formation of the House Range and Wah Wah Arch (Miller et al., 2003). Miller (1969) named and interpreted the Wah Wah Arch, the Tooele Arch and the House Range as half grabens. These originated above basement structures during the Late Cambrian (Poole, 1992). The basement structures consist of continental crust that rotated toward the spreading ridge separating the cratons of Siberia and Laurentia (Sears and Price, 2003).

1.10.1 Hellnmaria Member, Notch Peak Formation

The burrowed stromatolite facies in the Hellnmaria Member of the Notch Peak Formation is Upper Cambrian (Sunwaptan Stage) in age, and it is exposed on the Wah Wah Arch (Miller et al., 2012). Walcott (1908) named the Notch Peak Formation, and Hintze (1951) named the House Limestone. Hintze (1988) named the Hellnmaria, Red Tops, and Lava Dam members, and Miller (2001) revised the members and named members of the House Limestone. Miller *et al.* (2001) wrote a report that includes detailed descriptions of 17 measured sections totaling 2.98 km in thickness.

The Notch Peak Formation consists entirely of carbonates. The Hellnmaria

Member consists of dolomite and limestone (Figure 1.7), and stromatolites are abundant. All the samples collected and analyzed for this thesis are dolomitized.

The Hellnmaria Member in the northern part of the House Range is exposed as a sheer cliff face and is very difficult to access, but access to the member is easy in the southern House Range, where the strata are gently sloping along a minor dirt road (Figure 1.8) in Millard County, Utah.

1.11 South Lakeside Mountains

The South Lakeside Mountains are located along the southwest edge of the Great Salt Lake in Tooele and Box Elder Counties. The area is accessed from Exit 77 on Interstate 80. The North Lakeside Mountains are not accessible to the public due to military use. The Paleozoic stratigraphy in the South Lakeside Mountains was mapped by Young (1953), and the sedimentology of Upper Cambrian strata was described by Harlick (1989).

1.11.1 Upper Carbonate Member, St. Charles Formation

This section consists of Upper Cambrian limestone and dolomite dividing the Upper Cambrian section into two members of the St. Charles Formation: The Worm Creek Member and the Upper Carbonate Member. The Worm Creek Member is quartzite, and it occurs betweem the underlying Middle Cambrian Nounan Formation and the overlying Upper Carbonate Member. The St. Charles Formation is overlain by the Lower Ordovician Garden City Formation (Figure 1.9). The Garden City Formation is a laminated limestone lacking the stromatolites and burrows that typify the St. Charles

Formation.

During the Cambrian, sea level rose and inundated the continent, resulting in a time-transgressive sequence of quartzites overlain by siltstones and mudstones and then capped with carbonates (Dickinson, 1992; Stokes, 1986). The Upper Carbonate Member of the St. Charles represents the transgressive carbonates in this location in the South Lakeside Mountains. The Upper Carbonate Member in this area consists of blue-gray, bioturbated dolomite containing abundant stromatolites and burrows.

The stromatolites in the Upper Carbonate Member include laterally linked hemispheroids, stacked hemispheroids, and spheroidal structures. When viewed at the outcrop scale, the internal structure is preserved and there are distinct color differences between the stromatolites and the surrounding matrix. The microstructure is not well preserved due to extensive dolomitization. The burrows are usually lighter in color than the surrounding material.

Chert layers and nodules are common in some areas. This chert is dark orange and tan where it has been weathered and is black inside the nodules where it has not been weathered. The chert occurs as branching structures on bedding planes exhibiting "T" and "Y" branching angles that represent over-silicified burrow systems.

1.12 Regional Stratigraphy

The rocks in the Great Basin of Utah, where the strata in this study are exposed, are part of the Sauk Megasequence (Miller et al., 2011). Stratigraphic evidence of the Sauk Megasequence in the field areas begins with a basal Cambrian transgressive sandstone, called the Prospect Mountain Quartzite, overlain by Middle and Upper

Cambrian, mostly limestone and dolomite, overlain by deep-water shale intervals, and then capped by Ordovician sandstones. The Ordovician sandstones mark a regression that ends the Sauk Megasequence, and the corresponding rock unit is called the Watson Ranch Quartzite (Figure 1.10) (Miller et al., 2011).

Cambrian and Ordovician strata in Millard and Juab counties, western central Utah, are composed of a thick (5334 m) succession that was deposited on a tropical platform that experienced rapid subsidence (Miller et al., 2011). This subsidence created an ideal situation for complete deposition of regressive and transgressive sequences in the Great Basin. The Upper Cambrian highstand is exposed in the field area, and contains the burrowed stromatolitic facies found in both the Upper Carbonate Member of the St. Charles Formation, and in the Hellnmaria Member of the Notch Peak Formation.

Deposition of a thick succession of underlying strata combined with an abrupt rise of sea level may have triggered renewed slippage along the fault at the south edge of the House Range embayment. The facies patterns suggest that movement along this fault continued from deposition of the Middle Cambrian Wheeler Shale to the deposition of the Upper Cambrian Notch Peak Formation (Miller et al., 2011).

Subsidence in the Great Basin during Cambrian deposition was rapid enough that the strata are uncommonly thick and preserve an unusually complete depositional record. The Great Basin has a record of lowstand deposits, instead of the unconformities usually associated with lowstand deposits (Miller et al., 2003). This is one reason why the Red Tops Member lowstand, which is underlain by the Hellnmaria highstand, truncates the Hellnmaria stromatolitic facies and is readily discerned in the strata of the area.

1.12.1 House Range

In the House Range, the Notch Peak Formation consists entirely of carbonate strata and the Hellnmaria Member is dolomite and limestone, and stromatolites are common. Body fossils are rare in the dolomite facies. The Red Tops Member directly above the Hellnmaria Member is much thinner due to being deposited during a shoreline regression.

1.12.2 Hellnmaria Member

The cliffs of Notch Peak are almost vertical and consist almost entirely of the Hellnmaria Member with a small section of the Red Tops Member at the summit. The bedding is emphasized by thermal metamorphism (Miller et al., 2012) and gives the cliffs a light- and dark-gray striped pattern.

The Hellnmaria Member is informally divided into parasequences 1-4. The lower unit is a deep-water platy limestone, and then the facies shallow upward and end as peritidal facies near the top (Miller et al., 2011). This sequence was deposited during the Hellnmaria highstand, represented by parasequences 1 through 3. Sequence 4 is a widespread stromatolitic biostrome truncated by the Red Tops Member contact. Parasequence 4 is a thin light-gray interval above an irregular paleokarst surface and includes the top of the Hellnmaria Member. It is the unit that contains the burrowed stromatolites, where samples were collected for this research. The Hellnmaria Member is exposed in several areas including the collection site on the Wah Wah Arch near Steamboat Pass. At this location, the lower half of the Hellnmaria Member is massive limestone and the remainder is dolomite. Exposed at this location are rocks containing

ooids, oncoids, and stromatolites (Figure 1.11).

1.12.3 Red Tops Member

The Red Tops Member lies directly above the Hellnmaria Member. The Red Tops Member displays represents a relative sea level lowstand and is a sequence boundary between Sauk 4 and 5 (Miller et al., 2006). Stromatolites and burrows are not present in the Red Tops Member at Steamboat Pass in the bedding unit right above the Hellnmaria Member.

1.12.4 South Lakeside Mountains

The St. Charles Formation interfingers with the overlying Garden City Formation. The principal stratigraphic boundaries are placed at the top of the highest traceable bed of the underlying unit. The lower contact of the Garden City Formation is the Worm Creek Member and the Upper Carbonate Member of the St. Charles Formation. The Worm Creek Member consists of quartzite and is easy to identify in the field because it is a tan colored, coarse-grained, siliclastic rock that is heavily burrowed with *Skolithos* trace fossils. The differing trace fossils are also a diagnostic feature. The rock that overlies it is a blue-gray dolomite with stromatolite facies and burrow-like structures.

The contact between the St. Charles Formation (Upper Cambrian) and the overlying Garden City Formation (Lower Ordovician) defines the boundary between the Cambrian and the Ordovician in the South Lakeside Mountains. This is similar to the stratigraphy in the southern House Range. The stromatolite facies is in the Upper Carbonate Member. The Garden City Formation rocks show evidence of sea level

regression, which correlates to the Red Tops Member to the south.

1.13 The Upper Carbonate Member and the Hellnmaria Member

The Upper Carbonate Member is quite thick, like the Hellnmaria Member, and both units contain dolomitized, stromatolite facies with burrow-like structures. Changes in color and trace fossils mark the transition from the Upper Carbonate Member to the Garden City Formation. The Upper Carbonate Member is blue-gray dolomite, mostly burrowed, and the Garden City Formation is more orange or red in color, and it is laminated dolomite. The Red Tops Member in the southern House Range is also orange/red, and it is generally laminated with no visible trace fossils or stromatolites (Figure 1.12).

Cambrian seas transgressed onto Laurentia resulting in siliclastic deposits such as quartzite, which grades upward into carbonates due to the long, continuous rise of sea level through the Cambrian (Miller et al., 2011). This sequence of events is evident in both field areas. The resulting strata are a relatively thick Upper Cambrian, heavily burrowed, stromatolitic, stratigraphic succession along the western edge of the Great Basin.

1.14 Depositional Patterns in the Cambrian

The uppermost Cambrian in the southern House Range of Utah was deposited on the Wah Wah Arch, which was mostly a restricted marine depositional environment (Miller et al., 2011). The rocks on the Wah Wah Arch are thick, shallow marine carbonates. Most of the rock from the Upper Cambrian in the Hellnmaria Member of the

Notch Peak Formation in the House Range and the Upper Carbonate Member of the St.

Charles Formation in the South Lakeside Mountains is dolomite.

1.14.1 Paleobathymetry

Since the burrows observed in this study are far too old to have been created by

decapod crustaceans, the identity of the burrowers is unknown, thus eliminating the

possibility of inferring paleobathymetry using knowledge of the habitat requirements of a

specific kind of organism.

Trace fossil morphology is strongly controlled by the behavior of the producers,

but it is important to keep in mind that identical trace fossils may be the work of

unrelated animals (Seilacher, 2007; Buatois and Mangano, 2011). This may be the case

with the Late Cambrian burrows in this study. Comparisons of trace fossil bathymetry

can be used if the behavior of the organisms within their environment is taken into

account, including other factors that change with water depth as well (Ekdale, 1988).

Because trace fossils are generally autochthonous, it can be assumed that they are found

in the strata in the place where they formed, unless it is clear they are preserved in a

transported clast. *Thalassinoides* is a facies-breaking ichnogenus when it comes to

interpreting bathymetry, so if these are indeed *Thalassinoides,* other considerations must

be taken into account when determining a depositional environment. The burrows

structures found in this study are directly associated with the stromatolites, which provide

more reliable evidence of paleobathymetry.

Seilacher (1967) observed recurrent assemblages of trace fossil taxa having

paleoenvironmental implications, which led him to establish the concept of ichnofacies.

An ichnofacies is a characteristic association of trace fossils directly reflecting certain environmental conditions. Bathymetry, salinity, and substrate are examples of the environmental conditions which may be directly associated with ichnofacies (Ekdale, 1988). The *Glossifungites* ichnofacies represents infaunal communities in marine firmgrounds (i.e., stiff compacted muds), where mainly crustaceans and bivalves have dug their dwellings (Pemberton and Frey, 1985; Pemberton et al., 2004).

Organisms that have created trace fossils in marine environments are distributed according to three main factors: water depth, water chemistry, and substrate character. In marine settings as the water becomes deeper, there is a decrease in temperature, light, wave energy, sediment rate, sediment grain size, and food supply (Ekdale, 1988). In this way, the bathymetry of trace fossils is controlled by a multiplicity of environmental factors.

1.14.2 Evidence of a Shallow Marine, Restricted Platform

The Upper Cambrian strata in the Great Basin of Utah were deposited on an open marine platform with a platform edge lying to the west. The rocks contain trace fossil evidence showing disturbed bedding by burrowing organisms. The ichnofabric index (ii) in these rocks is mostly ii3 to ii5, indicating an abundance of burrowing organisms.

The depositional environment of the St. Charles Formation was a marine platform with platform-edge sands. The platforms included are restricted and open marine facies. The open marine platform had greater water circulation and contained more environmentally tolerant species of benthic animals, resulting in intense and extensive bioturbation. In modern environments, algal boundstones are restricted to hostile, often

hyper-saline environments. During the Upper Cambrian algal boundstones could extend to the oolitic platform-edge sands, because the paleoenvironment was more favorable for production of algal boundstones, including stromatolites.

Burrowing organisms thrive in open marine environments, because wave action provides an abundance of food material and oxygen. Most of the burrows in this study are associated with stromatolites. Water energy at the platform edge may have been low enough for stromatolites and ooids to form and not be ripped apart by larger waves or storm waves. The stromatolites were an important source for shelter and food for a variety of benthic organisms. The St. Charles Member and the Hellnmaria Member both contain a significant amount of algal boundstone as well as burrowed stromatolites.

There are modern analogies for carbonate deposition in places near intertidal inner shelf and outer shelf areas (Carannante et al., 1988). They are found around modern continents where siliclastic sediment input is low, such as the perimeter of Florida, or on isolated shallow marine banks, such as the Bahama Banks where the sedimentary succession under the platform is up to 8 km thick. Another example of a modern day carbonate platform is the Yucatan Peninsula, where the sedimentary succession is up to 2 km thick. These carbonate platforms are confined to tropical latitudes, have low sediment input, and are well within the photic zone (Carannante et al., 1988). The Upper Cambrian depositional environment of the Hellnmaria and Upper Carbonate members would have been similar to these modern carbonate platforms.

Chert is uncommon in most of the Upper Cambrian sections, but there are locally abundant nodules of chert in both field areas and an abundance of silicified burrows. The original silica source for the chert may have been from sponge spicules. Siliceous

sponges are known from this time, although no undoubted sponge fossils have been discovered at the two field areas for this project.

The depositional environment conductive for stromatolite growth would need to be a setting with little or no terriginous influx, as too much sediment in the water column would inhibit photosynthesis needed for algal growth. To keep the water sufficiently oxygenated to support animal life, there would need to be some wave action, but the water energy would have to be low enough to prevent the stromatolites from breaking apart. The most likely scenario to fit all these conditions would be a shallow subtidal zone within fair weather wave base in an open marine setting.

Landward of the outer platform edge, open lagoons and bays could have created an environment for environmentally tolerant species, such as the stromatolites that we find in the Upper Cambrian fossil record. Lagoons and tidal flats in this type of environment undergo alternating periods of exposure and then flooding. This results in a stressful environment for normal animal life, whereas tolerant primary producers such as stromatolitic cyanobacteria can thrive.

The oncoids observed in the two study areas are concentrically laminated, indicating an environment of oscillating motion during formation. Concentric oncoids are restricted underwater areas with enough agitation to cause continued motion (Bathhurst, 1975). Most of the oncoids observed in the study areas occur in lime packstones.

Interbedded lime mudstone and grainstone is common in the Upper Carbonate Member as well as in the Hellnmaria Member. Usually, lithofacies such as these are deposited in a lagoonal environment (Wilson, 1975). Interpreting the depositional environment of Cambrian rocks can be difficult because of the lack of direct modern

analogues. Some of the Cambrian paleoenvironments are unknown in our present-day oceans, because today there is a larger percentage of water fixed in the form of ice at the poles, resulting in lower sea levels today. Another problem is that Cambrian marine organisms were quite different from modern marine organisms, and many of the Cambrian animals were soft-bodied invertebrates with a poor to nonexistent fossil record. All these factors contribute to making the principal of uniformitarianism more difficult to apply to Late Cambrian ecosystems.

Several different types of algal boundstones, such as stromatolites and oncolites, in the both the Hellnmaria Member and the Upper Carbonate Member, are restricted to hostile environments where they are emergent for part of the time. During the Late Cambrian, it has been suggested that due to a lack of grazing organisms, stromatolites thrived in less hostile environments, such as those suggested in these strata (Shapiro, 2007). Gastropod and polyplacophore body fossils are found in association with the Upper Cambrian stromatolites at both Utah localities (Vendrasco and Runnegar, 2004), and these mollusks are representative of a low-energy, shallow marine environment.

Carbonate shelf environments are not always laterally continuous for large distances, adding to their complex nature. Tidal flats and lagoons may contain mudstone and grainstone sediments occurring side by side. Biology is a major factor on carbonate deposition, so the organisms play a crucial role in determining what types of rocks are formed.

<u>1.15 Global Sea Level Changes</u>

Transgressive sandstones mark the base of the Cambrian in western Utah and throughout the Great Basin. As the shoreline moved eastward and the sea grew deeper, deposits of fine-grained, silty clay and mud became shale, mudstone, and siltstone of the Early and Middle Cambrian (Burchfiel, 1992). As the transgressive sequence continued into the Late Cambrian, deposits of limestone began to form as a result of biological activity (Burchfiel, 1992; Miller et al., 2011). The Hellnmaria Member and the Upper Carbonate Member are both Late Cambrian limestone and dolomite units (Dickinson, 1992; Hintze and Kowalis, 2009). They were deposited during a sea level highstand, and the rock formations that lie just above both members show evidence of a regression at the end of the Cambrian. The Hellnmaria Member rocks shows a record of high sea level known as the Hellnmaria highstand (Miller et al., 2011).This is consistent with deposition during a greenhouse sea pattern without polar icecaps (Dickinson, 1992). The overlying Red Tops Member records a sea level lowstand and truncates the Hellnmaria Member in the southern House Range.

During the Late Cambrian in Utah, when the Hellnmaria and Upper Carbonate Members were deposited, the area was a carbonate platform with extensive limestone deposition (Dickinson, 1992). Under natural conditions, water pressure may be lowered and more carbon dioxide absorbed by the atmosphere by the breaking of waves in the surf zone or over shallow banks. This supports the interpretation that the depositional environment in the study areas was in shallow marine, alkaline, warm tropical waters.

The ocean may have been more alkaline than normal, but the difference may have been small due to the fact that the ocean generally is a well-buffered solution. The actual

change in pH in seawater caused by the gain or loss of carbon dioxide is quite small, and values rarely fall outside of 7.8 to 8.3 (Bathhurst, 1975). Alkalinity may have increased during the Late Cambrian due to organic material matter degradation through sulfate reduction by bacteria (Bernasconi, 1994).

Rapid sea floor spreading correlates with high sea levels during the Upper Cambrian and the carbon dioxide levels were higher in the atmosphere during this time because of the increased rate of carbon dioxide outgassing related to high rates of sea floor spreading. This also leads to higher atmospheric temperatures that may have facilitated limestone deposition during the Late Cambrian. Strict uniformitarianism is problematic once again, because we do not have exactly the same conditions existing anywhere on Earth today.
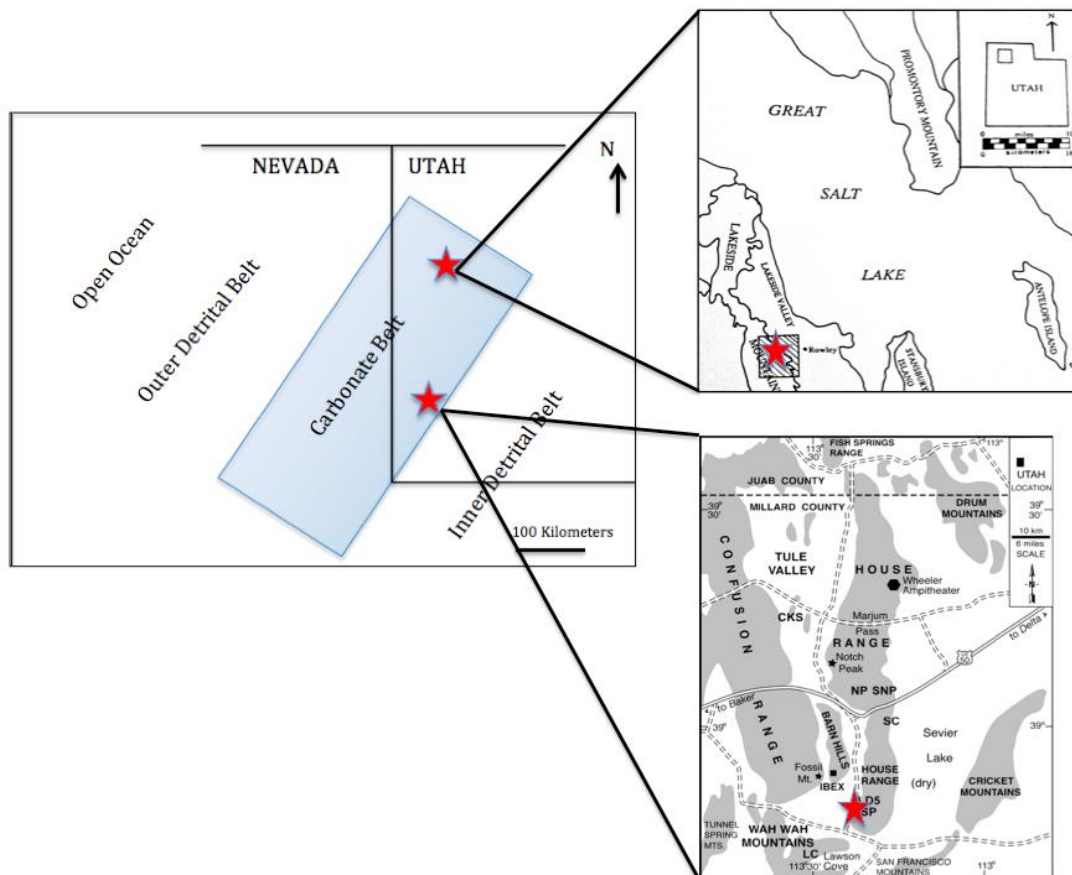
Figure 1.1. Generalized map showing the location of the two field areas. The red star to the North is the South Lakeside Mountains location, and the red star to the south is the southern House Range location. Both field areas were situated within the middle carbonate belt in the Late Cambrian, which was a tropical, low energy, shallow marine environment during this time (modified after Palmer, 1971); and (modified after Miller, 2011).
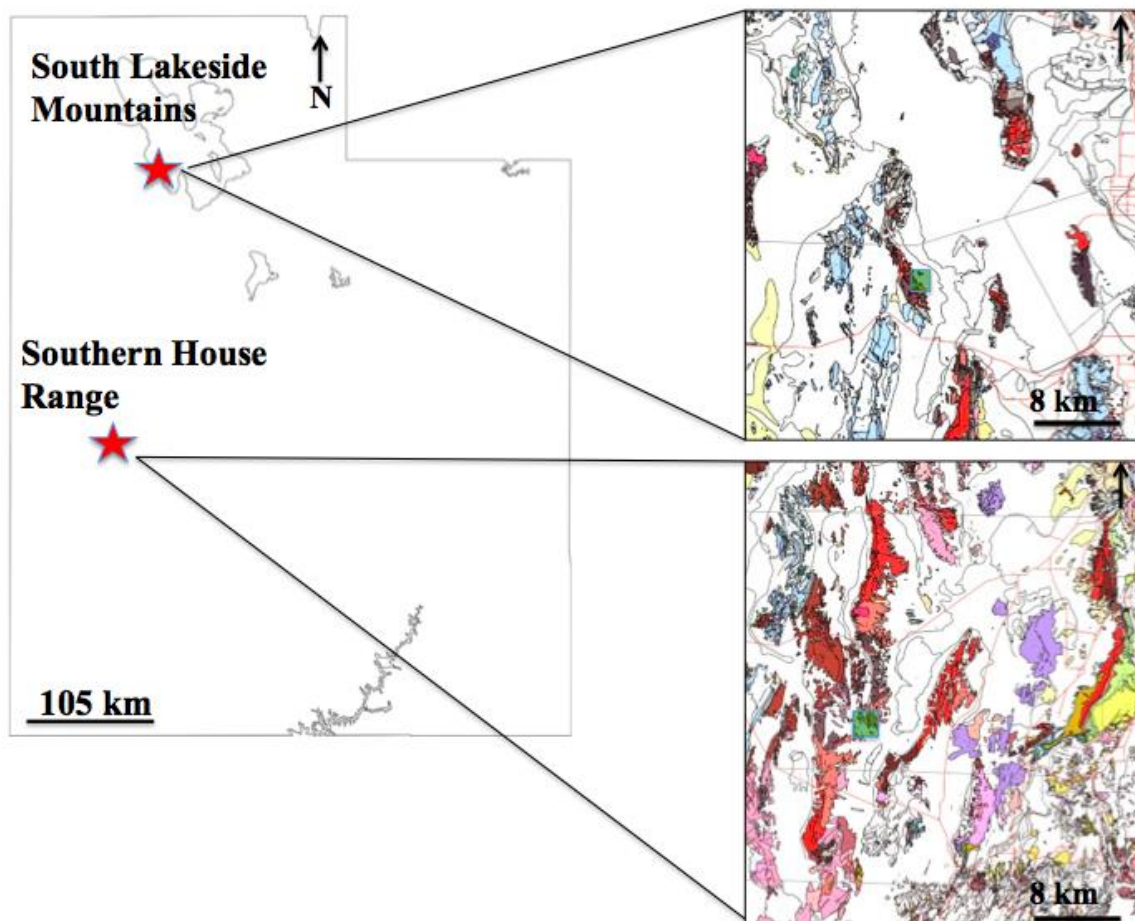
Figure 1.2 Details of geologic maps of the two field areas. Red stars show where samples were collected and observed in both field areas. The geologic maps have a green square showing field sites. The light orange color on the maps denotes the Late Cambrian rocks (modified from Hintze et al., 2000).
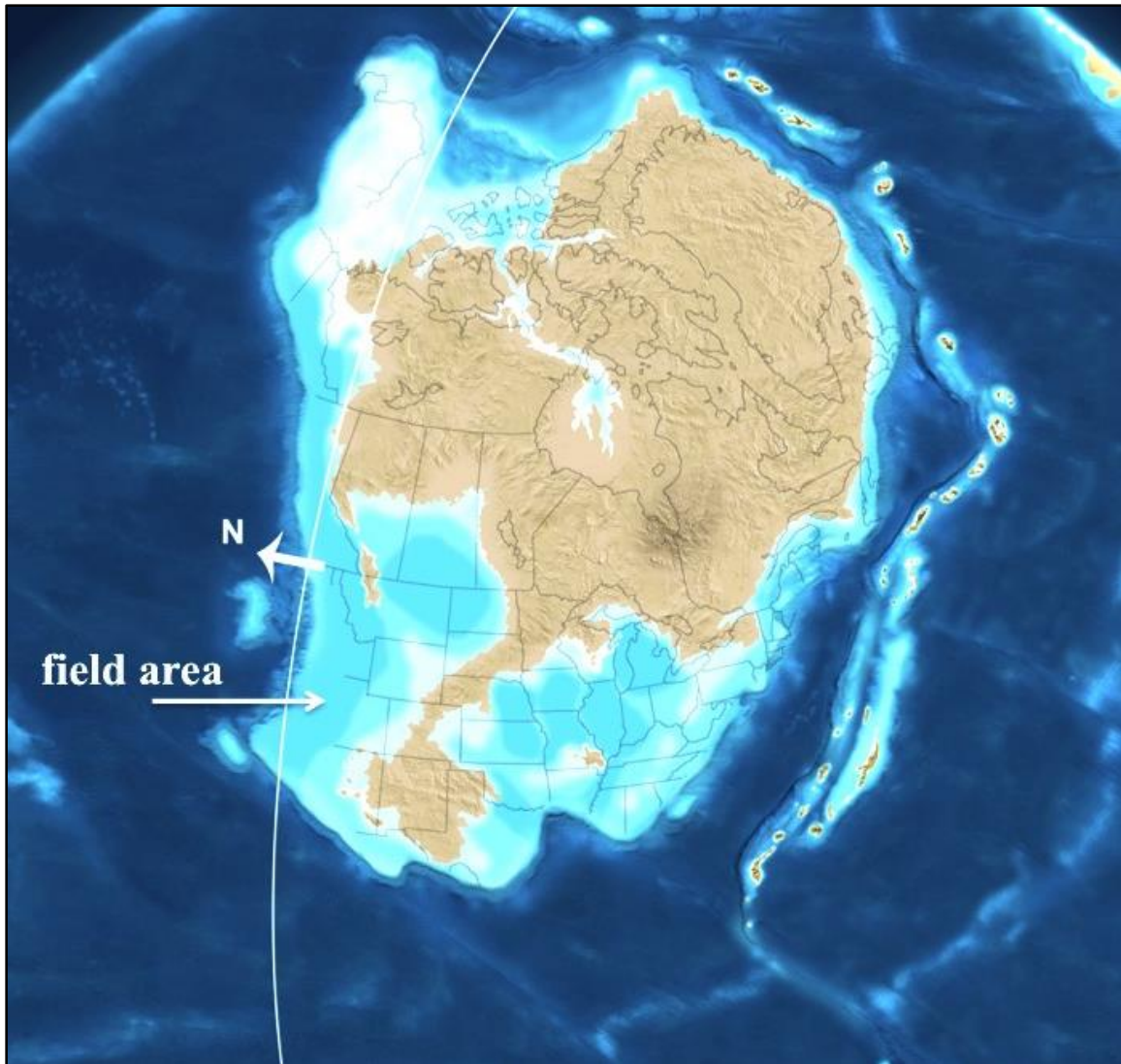
Figure 1.3 Paleogeographic map of North America during the Late Cambrian, including the distribution of ancient shallow seas, deep ocean basins, mountain ranges, coastal plains, and continental interiors. The white arrow marks the approximate location of the study sites, which are located in a passive marine margin setting (modified from Blakey, 1996).

Figure 1.4 Example of a burrowed stromatolite found in the Late Cambrian middle carbonate belt carbonate belt along the eastern part of the Great Basin. Plan view.
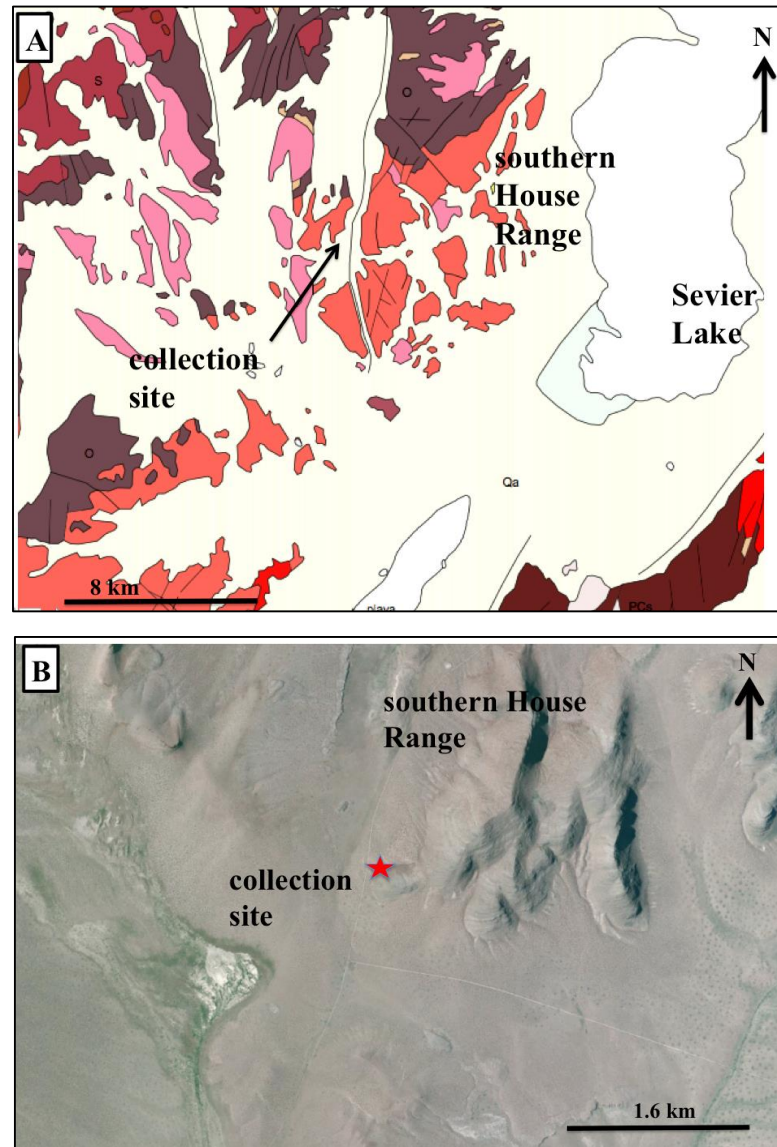
Figure 1.5. Map and photo of primary field area.(A) Digital geologic map of the southern House Range, Utah. The orange colored sections are Late Cambrian exposures. The arrow shows the field area of the southern House Range where the stromatolitic burrowed facies are exposed and easy to access (modified from Hintze et al., 2000). (B) Google Maps image of the Steamboat Pass locality in the southern House Range. The red star shows where specimens were collected.
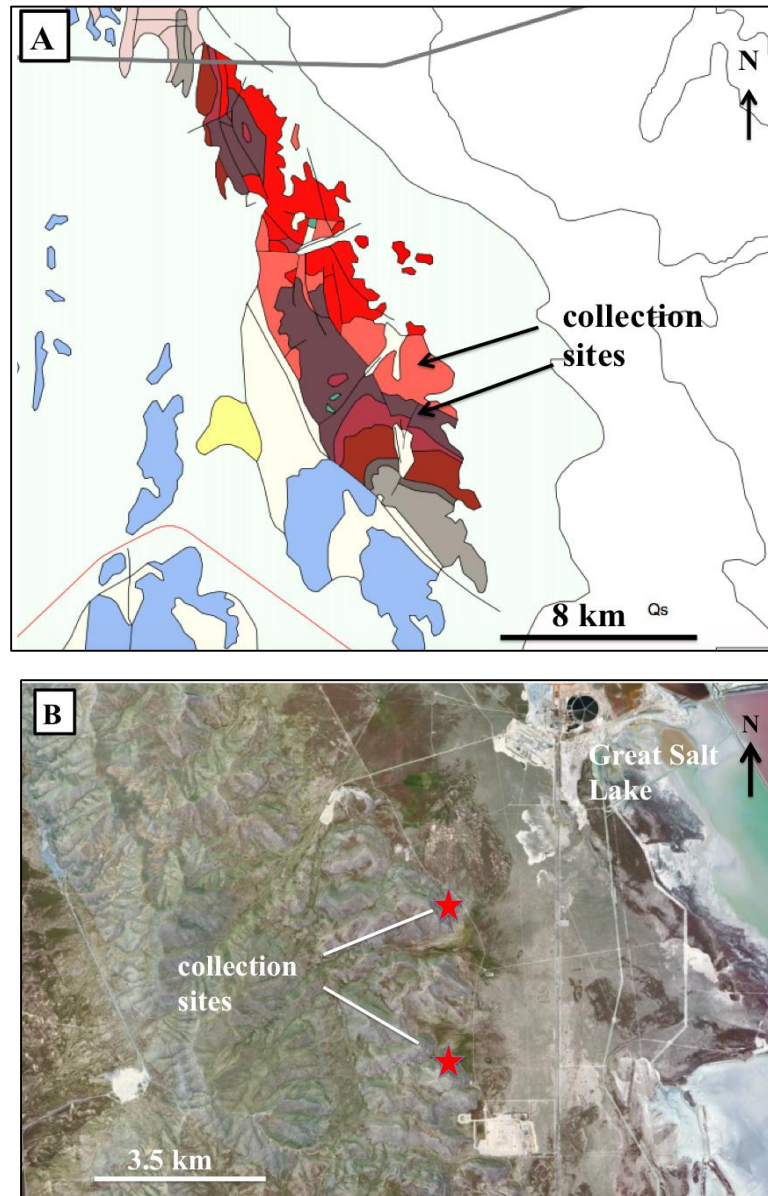
Figure 1.6 Map and photo of field area with correlating strata to the primary field area. (A) Digital geologic map of the South Lakeside Mountains, Utah. The orange colored sections are Late Cambrian exposures. The black arrows show the exposed stromatolitic facies (modified from Hintze et al., 2000), (B) Google Maps image of the same area as in figure A.

L. ORDOVICIAN	Tremadocian	FILLMORE FORMATION (550 m)	HOUSE LIMESTONE
		Red Canyon Member (69.8 m)	
		Burnout Canyon Member (19.5 m)	
		Barn Canyon Member (75.6 m)	
FURONGIAN SERIES	Lawsonian Stage	Lava Dam Member (76.5 m)	NOTCH PEAK FORMATION
		Red Tops Member (26.5 m)	
	Jiangshanian Stage	Hellnmaria Member (318 m)	
	Paibian	Sneakover Limestone Member (50 m)	ORR FORMATION
		Steamboat Pass Shale Member (80 m)	
SERIES 3	Stage 7	Big Horse Limestone Member (210 m)	

Figure 1.7. The blue shaded area marks the section where samples in this study were collected and analyzed. This figure shows that the Hellnmaria Member is a thick unit overlain by the Red Tops Member (modified after Landing, 2011).

Figure 1.8 Notch Peak, located north of the study area at Steamboat Pass. The nearly vertical cliffs of Notch Peak consist almost entirely of the Hellnmaria Member with a small section of the Red Tops Member at the summit. It has been suggested that the striped nature of the bedding in the Hellnmaria Member reflects periodic cycles (Miller, 2012). Just below the Red Tops Member is correlative strata to the collection site to the south of where this photo was taken.
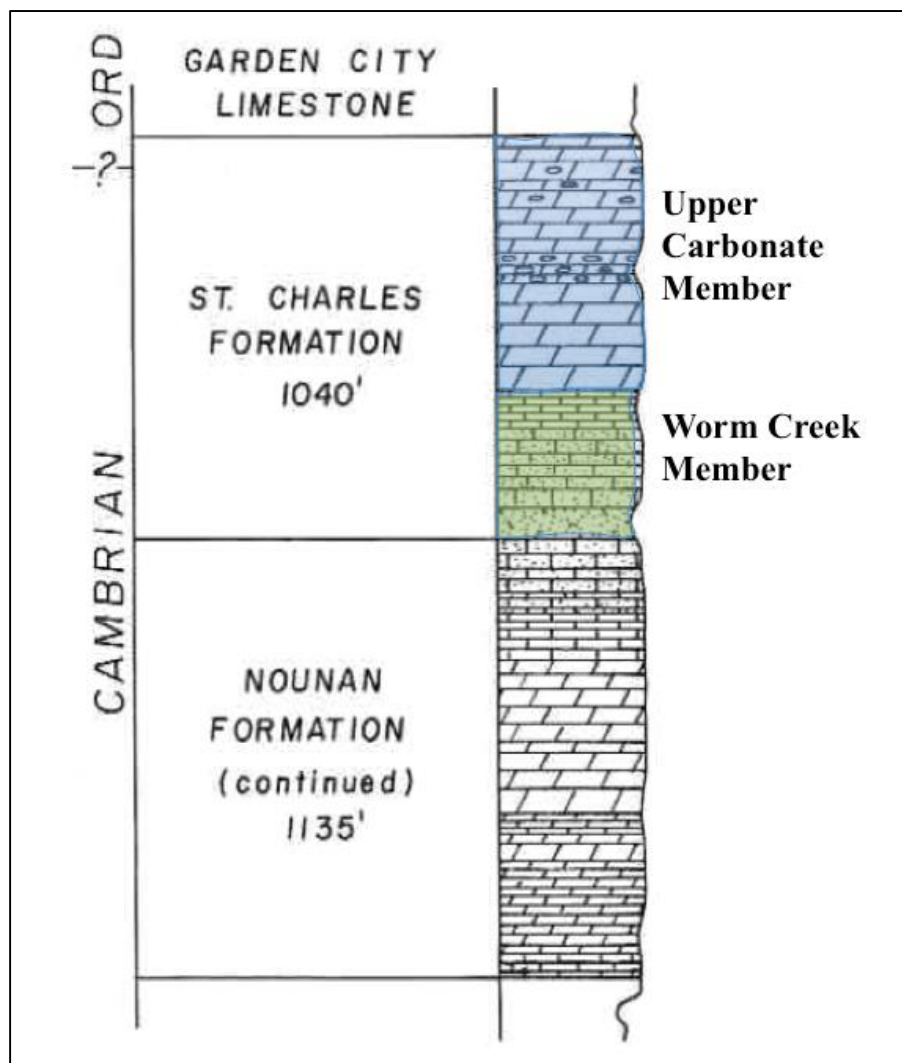
Figure 1.9 Stratigraphic column of Upper Cambrian strata in the South Lakeside Mountains. The Upper Carbonate Member is the section where samples were observed and collected (modified after Stokes, 1986).
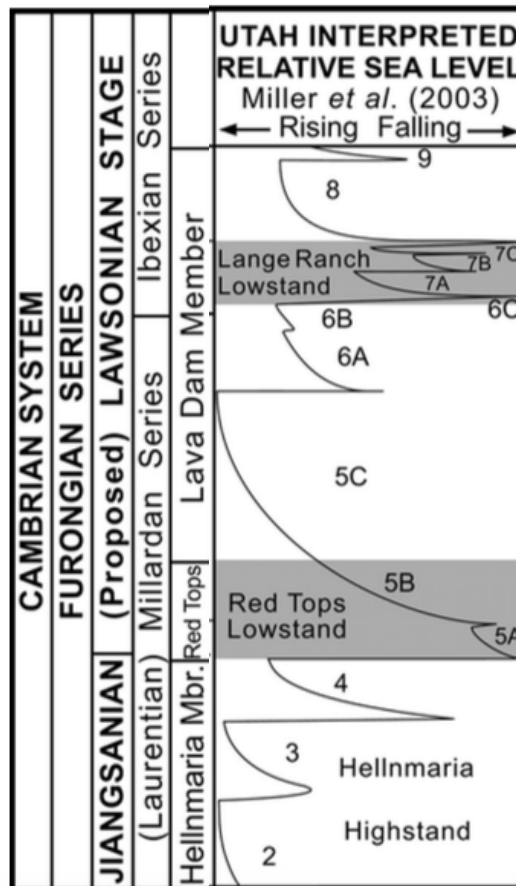
Figure 1.10 The Hellnmaria Highstand in this figure is the approximate stratigraphic position where the samples were collected. This figures show that the Hellnmaria and Upper Carbonate Members were deposited during a sea level highstand, and the overlying Red Topps Member was deposited during a sea level lowstand. When the sea level was high, the facies were stromatolitic, but the stromatolites largely disappeared as the sea level dropped (modified after Miller et al., 2011).
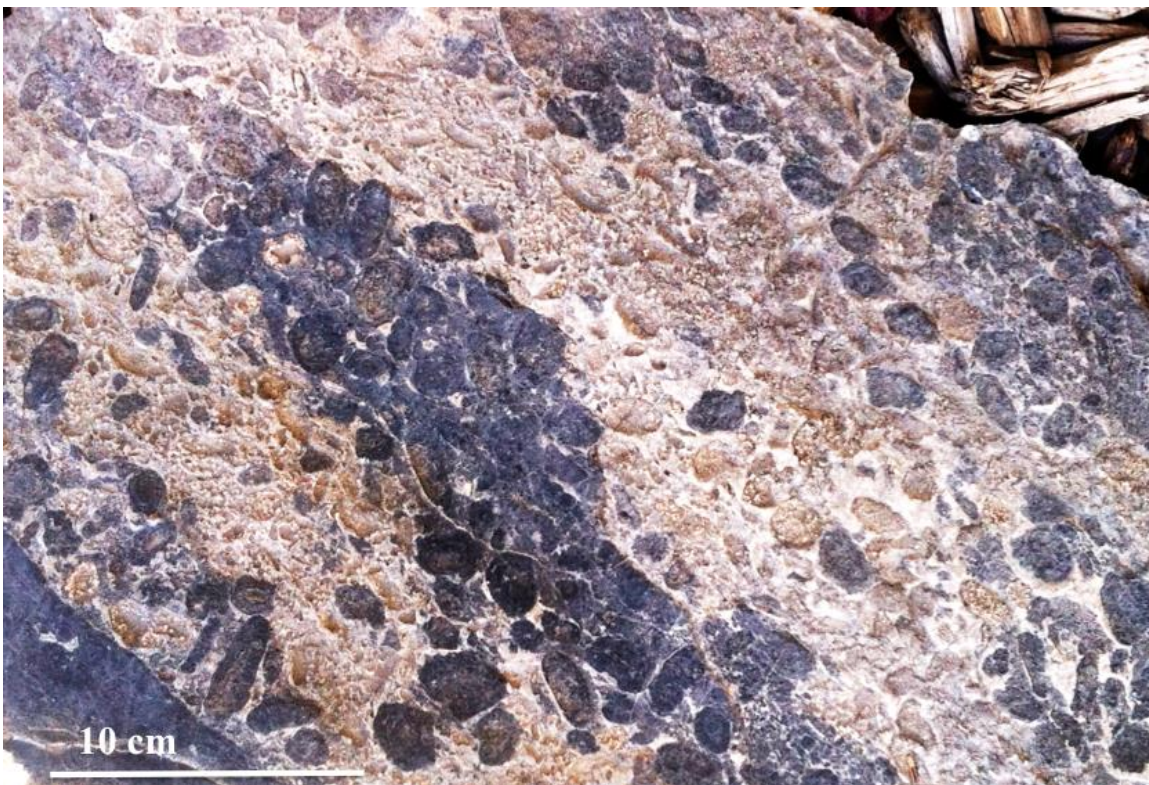
Figure 1.11 Concentrically laminated oncoids deposited in a layered pattern just below the burrowed stromatolites in the Hellnmaria Member of the Notch Peak Formation in the southern House Range. This oncloidal layer forms the base from where many stromatolites were initiated. This same pattern is found in the South Lakeside Mountain field site.
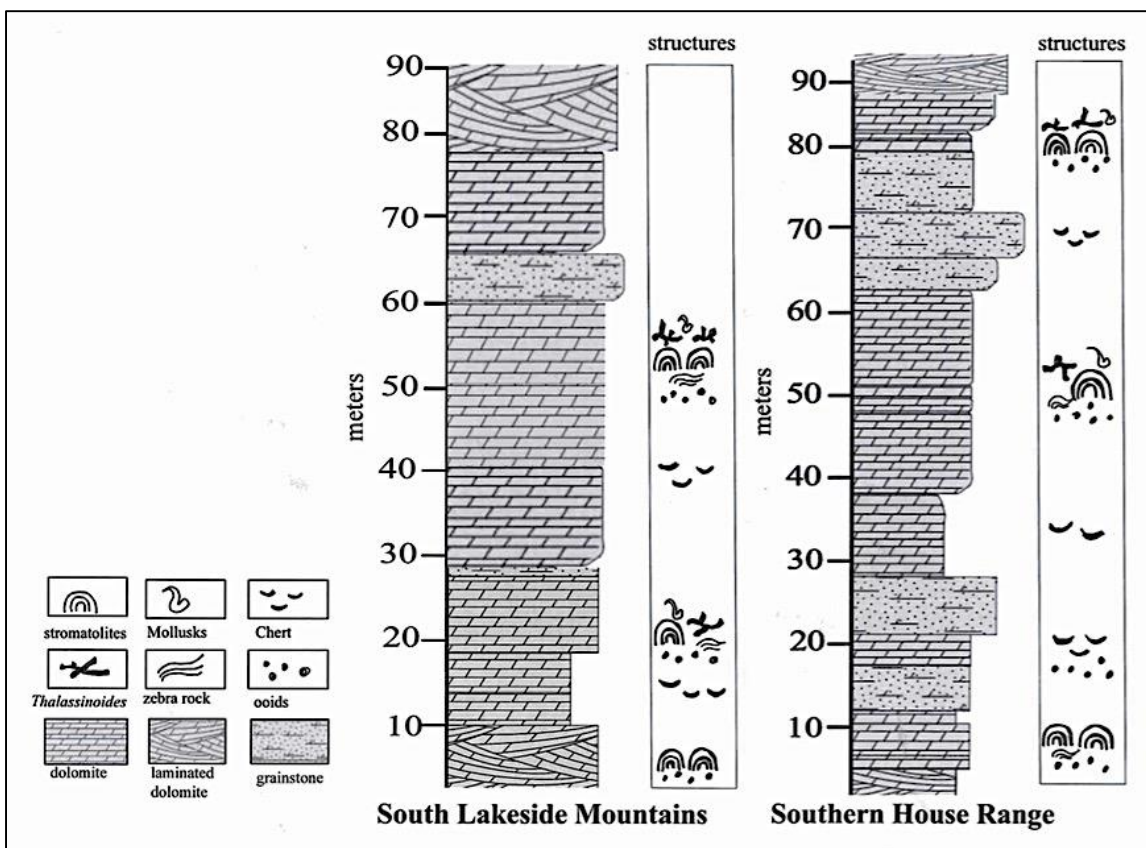
Figure 1.12 Stratigraphic columns in both field areas showing the sedimentary structures founds in field areas (measured in meters).

1.16. References

ALROY, J., D. J. BOTTJER, M. FOOT, F. T. FURSICH, P. J. HARRIES, A. J. W. HENDY, S. M. HOLLAND, L. C. IVANY, W. KIESSLING, M. A. KOSNIK, C. R. MARSHAL, A. J. MCGOWAN, A. I. MILLER, T. D. OLSZEWSKI, M. E. PATZKOWSKY, S. E. PETERS, L. VILLIER, P. J. WAGNER, N. BONUSO, P. S. BORKOW, B. BRENNEIS, M. E. CLAPHAM, L. M. FALL, C. A. FERFUSON, V. L. HANSON, A. Z. KRUG, K. M. LAYOU, E. H. LECKEY, S. NURNBERG, C. M. POWERS, J. A. SESSA, C. SIMPSON, A. TOMAOVCH AND C. C. VISAGGI. 2008. Phanerozoic trends in the global diversity of marine invertebrates. Science, 321:97–100.

AWRAMIK, S. M. 1971. Precambrian columnar stromatolite diversity: reflection of metazoan appearance. Science, 174:825–827.

BATHHURST, R. G. C. 1975. Carbonate Sediments and Their Diagenesis (second edition). Developments in Sedimentology 12. Amsterdam, Elsevier 459–473 p.

BERNASCONI, S. M. 1994. Geochemical and microbial controls on dolomite formation in anoxic environments: A case study from the Middle Triassic (Ticino, Switzerland). Contributions to Sedimentology 19, Schweizerbartsche Verlagsbuchhandlung, Stuttgart, 109 p.

BERNER, R. A. 2003. The long-term carbon cycle, fossil fuels and atmospheric composition. Nature, 426:323–326.

BERNER, R. A. 2009. Phanerozoic atmospheric oxygen: New results using the GEOCARBSULF model. American Journal of Science, 309:603–606.

BERNER, R. A. AND Z. KOTHAVALA. 2001. GEOCARB III: A revised model of atmospheric CO_2 over Phanerozoic time. American Journal of Science, 301:182−204.

BUATOIS, L. AND M. G. MANGANO. 2011. Ichnology: organism-substrate interactions in space and time. New York, Cambridge University Press, 358 p.

BURCHFIEL, B. C., D. S. COWEN, AND G. A. DAVIS. 1992. Tectonic overview of the Cordilleran orogeny in the western United States, p.417–422. *In* B. C. Burchfiel, P. W. Lipman, and M. L. Zoback, (eds.), The Cordilleran Orogen: Conterminous U.S.: Boulder, Colorado, Geological Society of America. The Geology of North America, G-3.

BLAKEY, R. C. 1996. Permian eolian deposits, sequences, and sequence boundaries, Colorado Plateau, p. 405−426. *In* M. W. Longman and M. D. Sonnenfeld, (eds.), Paleozoic systems of the Rocky Mountain Region: Rocky Mountain Section. Society of Sedimentary Geology.

CARANNANTE G., M. ESTEBAN, J. D. MILLIMAN, AND L. SIMONE. 1988. Carbonate lithofacies as paleolatitude indicators: problems and limitations. Sedimentary Geology, 60:333−346.

COOK, H. E. 1992. Comparison of Paleozoic passive-margin carbonate platforms of the western United States and the USSR. Houston Geological Society Bulletin, 34:19.

DICKINSON, W. R. 1992. Cordilleran sedimentary assemblages, p.417–422. *In* B. C. Burchfiel, P. W. Lipman, and M. L. Zoback, (eds.), The Cordilleran Orogen: Conterminous U.S.: Boulder, Colorado, Geological Society of America. The Geology of North America, G-3.

EKDALE, A. A. 1988. Pitfalls of paleobathymetric interpretations based on trace fossil assemblages. PALAIOS, 3:464−472.

FELDMAN, R. M. AND J. A. MCKENZIE. 1998. Stromatolite-thrombolite associations in a modern environment, Lee Stocking Island, Bahamas. PALAIOS, 13:210212.

FELDMANN, R. M. AND C. E. SCHWEITZER. 2010. The oldest shrimp (Devonian: Famennian) and remarkable preservation of soft tissue. Journal of Crustacean Biology, 30:629−635.

GRADSTEIN, F. M., J. G. OGG and M. D. SCHMITZ, G. M. OGG. 2012. The Geologic Time Scale 2012. Elsevier, Boston, USA, Elsevier, 1,144 p.

HARLICK, A. J. 1989. Sedimentary, paleoecology and diagenesis of the Upper Cambrian St. Charles Formation, Southern Lakeside Mountains, Utah. Unpublished M.S. thesis, University of Utah, Salt Lake City, Utah.

HINTZE, L. F. 1988. Geologic History of Utah. Brigham Young University Geology Studies, Special Publication, 7:202.

HINTZE, L. F., G. C. WILLIS, D. Y. M. LAES, D. A. SPRINKEL AND K. D. BROWN. 2000. Digital Geologic Map of Utah. Utah Geologic Survey.

HINTZE, L. F. AND B. J. KOWALIS. 2009. Geologic History of Utah. Brigham Young University Geologic Studies, Special Publication 9, 55−60 p.

KEPPER, J. C. 1981. Ramp and platform deposition during the Middle Cambrian, eastern California and western Nevada, p. 102-105. *In* M. E. Taylor (ed.), Short papers for the 2nd International Symposium on the Cambrian system, United States

Geological Survey Open File Report.

MILLER, J. F., K. R. EVANS, J. LOCH, J. D. ETHINGTON AND J. H. STITT. 2001. New lithostratigraphic units in the Notch Peak and House Formations (Cambrian-Ordovician), Ibex Area, western Millard County, Utah. Brigham Young University Geology studies, 46:35−69.

MILLER, J. F., R. L. ETHINGTON, R. E. EVANS, L. E. HOLMER, J. D. LOCH, L. E. POPOV, J. E. REPETSKI, R. L. RIPPERDAN AND J. F. TAYLOR. 2006. Proposed stratotype for the base of the highest Cambrian stage at the first appearance datum of *Cordylodus andresi*, Lawson Cove section, Utah, USA. Palaeoworld, 15:38−44.

MILLER, J.F., K. R. EVANS, R. L. FREEMAN, R. L. RIPPERDAN AND J. F. TAYLOR. 2011. Proposed stratotype for the base of the Lawsonian Stage (Cambrian Stage 10) at the First Appearance Datum of *Eoconodontus notchpeakensis* (Miller) in the House Range, Utah, USA. Bulletin of Geosciences, 86:595−620.

MILLER, J. F., K. R. EVANS AND B. F. DATTILO. 2012. The great American carbonate bank in the miogeocline of western central Utah: Tectonic influences on sedimentation, p.769−854. *In* J. R. Derby, R. D. Fritz, S. A. Longacre, W. A. Morgan, and C. A. Sternbach (eds.), The great American carbonate bank:The geology and economic resources of the Cambrian−Ordovician Sauk megasequence of Laurentia, American Association of Petroleum Geologists Memoir 98.

PALMER, R. T. 1971. The Cambrian of the Great Basin and adjacent areas, western United States: Cambrian of the New World, p. 79−80. *In* C. H. Holland (ed.), Wiley-Interscience.

PEMBERTON, S. G. AND R. W. FREY. 1985. The *Glossifungites* Ichnofacies: modern examples from the Georgia coast, USA, p. 237-259. *In* H. A. Curran (ed.), Biogenic structures: their use in interpreting depositional environments: Tulsa, Oklahoma, SEPM Special Publication 35.

PEMBERTON, S. G., MACEACHERN, J. A. AND T. SAUNDERS. 2004. Stratigraphic applications of substrate-specific ichnofacies: delineating discontinuities in the fossil record, p.29-62. *In* D. McIlroy (ed.), The application of ichnology to palaeoenviromental and stratigraphic analysis: Geological Society of London Special Publication 228.

POOLE, F. G. 1992. Latest Precambrian to latest Devonian time; Development of a continental margin, p.46. *In* B. C. Burchfiel, P. W. Lipman, and M. L. Zoback, (eds.), The Cordilleran Orogen: Conterminous U.S.: Boulder, Colorado, Geological Society of America. The Geology of North America, G-3.

SEARS, J. W. AND R. A. PRICE. 2003. Tightening the Siberian connection to western

Laurentia. Geological Society of America Bulletin, 115:9439−53.

SCOTESE, C. R. 1997. Continental Drift (7th edition.). Arlington, Texas: Paleomap Project, 79 p.

SEILACHER, A. 1967. Bathymetry of Trace Fossils: Marine Geology, 5:413−428.

SEILACHER, A. 1982. Fossil behavior. The Fossil Record and Evolution. Scientific American, 217:72−80.

SEILACHER, A. 2007. Trace Fossil Analysis. Berlin, Springer-Verlag, 226p.

SHAPIRO, R. S. 2007. Stromatolites: A 3.5 Billion-Year Ichnologic Record, p. 382–390. In W. Miller III (ed.), Trace Fossils: Concepts, Problems, Prospects. Amsterdam, Elsevier.

STANLEY, S. M. AND HARDIE L. A. 1999. Hypercalcification: Paleontology Links Plate Tectonics and Geochemistry to Sedimentology. GSA Today, 9(2):1–7.

STEWART, J. H. 1992. Latest Precambrian to latest Devonian time; Development of a continental margin, p.16−19. In B. C. Burchfiel, P. W. Lipman, and M. L. Zoback, (eds.), The Cordilleran Orogen: Conterminous U.S.: Boulder, Colorado, Geological Society of America. The Geology of North America, G-3.

STOKES, W. L. 1986. Geology of Utah. Utah Museum of Natural History, University of Utah and Utah Geologic and Mineral Survey, Department of Natural Resources, Salt Lake City, Utah, 47 p.

TAYLOR, M. E., AND H. E. COOK. 1976. Continental shelf and slope facies in the Upper Cambrian and Lower Ordovician of Nevada. Brigham Young University Geology Studies, 23(2):181–214.

VENDRASCO, M. J. AND B. RUNNEGAR. 2004. Late Cambrian and Early Ordovician stem group chitons (Mollusca: Polyplacophora) from Utah and Missouri. Journal of Paleontology, 78(4):675–689.

WILSON, J. L. 1975. Carbonate Facies in Geologic History. Springer-Verlag, New York, 469 p.

YOUNG, J. C. 1953. Geology of the southern Lakeside Mountains. Unpublished Master of Science Thesis, University of Utah, 90 p.

ZIEGLER, A. M., C. R. SCOTESE, W. S. MCKERROW, M. E. JOHNSON AND R. K. BAMBACH. 1979. Paleozoic paleogeography. Annual Review of Earth and Planetary Science, 7:473–502.

CHAPTER 2

ICHNOLOGY IN THE UPPER CAMBRIAN CARBONATE UNITS, HOUSE

RANGE AND SOUTH LAKESIDE MOUNTAINS, UTAH

2.1 Abstract

Upper Cambrian strata in the Great Basin of western Utah are dominated by

marine carbonate facies. Laminated microbial stromatolite facies were widespread during

the Cambrian-Ordovician transition in the eastern Great Basin of western Utah. Two

research areas, the South Lakeside Mountains and the southern House Range, both

located in the Basin and Range Geologic Province in western Utah, contain correlative

Upper Cambrian strata and burrowed stromatolite facies. The stromatolites contain a

heterogeneous sedimentary fabric exemplified by structures interpreted variously as

burrows (trace fossils) or inorganic diagenetic structures. This study concludes that they

are definitely burrows.

The Hellnmaria Member of the Notch Peak Formation in the southern House

Range and the Upper Carbonate Member of the St. Charles Formation in the northern

South Lakeside Mountains expose correlating cyanobacterial stromatolite facies

containing burrows. These shallow marine carbonate members carbonate units contain

abundant stromatolites, including laterally linked hemispheroids, stacked hemispheroids,

and columnar structures.

The Hellnmaria Member and the Upper Carbonate Member contain smooth-walled cylindrical features inside and outside individual stromatolites. They appear to be burrows, because they exhibit a branching morphology and have a coarser grain fill differing in texture from surrounding sediments. This study reveals that these Late Cambrian structures inside the stromatolites indeed are burrows, and because the branching geometry of the burrow tunnels is irregular and constricted, it is determined that they are the ichnospecies *Thalassinoides paradoxicus.*

Several burrows have been over-silicified because of growth of microcrystalline chert nodules extending beyond the boundaries of the original trace fossil. This is probably a result of favorable nucleation sites on the original burrow surface forming a nucleation site until the system runs low on silica.

2.2 Introduction

The texture, composition, and structure of sediment can be altered by the behaviors of organisms that inhabit the substrate. Ichnology focuses mainly on animal-sediment relationships and how to recognize and interpret these sedimentary structures, so ichnology is really the study of the fossil record of organism behavior (Seilacher, 1967b, 2007).

The probable Late Cambrian burrowers in this study likely were mining shallow marine muds for food as well as seeking protection in the complex web of burrows. Burrowing organisms of unknown identity constructed deep infaunal deposit feeding trace fossils (fodinichnia). These burrows are passively filled, and the different color and

texture of the burrows from the surrounding sediment suggests that they were open burrows that were later filled with sediments. *Thalassinoides* in the Late Cambrian are far too old to have been created by decapod crustaceans, which are the primary creators of such trace fossils in the Mesozoic and Cenozoic, so the identity of the Cambrian trace makers remains unknown. No body fossils were found that could be attributed to the trace makers, so soft-bodied organisms probably constructed them.

To demonstrate that these structures are burrow systems, and that the burrowers were sediment-ingesting deposit feeders as opposed to filter feeders or surface herbivores, approximately 50 samples of the stromatolitic facies from field areas in the South Lakeside Mountains and House Range of western Utah were collected for detailed study the lab. Serial sectioning of block specimens was performed on 10 of these samples. Burrowed stromatolite samples were sectioned both parallel and perpendicular to bedding planes in order to better define their three-dimensional network structure. These serial sections were polished to make morphological comparisons with known *Thalassinoides* burrows from other parts of the geologic record and to enhance the contrast between burrows and matrix. The average burrow angle was compared to other documented *Thalassinoides*.

Cambrian *Thalassinoides* were first reported by Myrow (1995). A majority of early Paleozoic *Thalassinoides* are commonly less than 10mm in diameter, architecturally simple, and are two-dimensional networks. There are larger 3-dimensional *Thalassinoides* described in more recent publications (Mángano and Buatois, 2007). *Thalassinoides* with a three-dimensional morphology has been described in Late Cambrian, Tremadocian carbonates in the Precordillera of Argentina (Canas, 1995).

2.3 Study Location

Both study sites are located in the Great Basin of western Utah, and both contain the stromatolite facies with burrow-like structures. The southern House Range is located southwest of Sevier Lake in Millard County (Figure 1.6). The South Lakeside Mountains are located along the southwestern side of the Great Salt Lake in Tooele County (Figure 1.7). All specimens from both study sites are from the Upper Cambrian strata (Figure 2.1) containing a stromatolitic facies with trace fossils. In the House Range, all specimens are from the Hellnmaria Member, Sunwaptan North American Regional Stage, of the Notch Peak Formation. Those collected from the South Lakeside Mountains are from the Upper Carbonate Member of the St. Charles Formation.

During the Late Cambrian, a north-trending carbonate belt extended along the continental margin in what is now Utah (Sears and Price, 2003). There are deposits representing the continental margin, continental slope, and continental rise sediments in eastern Utah continuing into deeper water to the west in Nevada (Dickinson, 1992; Taylor and Cook, 1976).

The Hellnmaria and Upper Carbonate Members were deposited in this carbonate belt during the Late Cambrian, in a shallow, tropical, quiet water paleoenvironment, perhaps in a highly alkaline setting on an open marine platform with a platform edge lying to the west (Robinson, 1960; Saltzman, 2005). The platform edge may have kept the energy low enough for stromatolites and ooids to form and not be ripped apart by larger waves or storm waves. The burrows that are preserved in both study areas exhibit a close association with stromatolites.

2.4 Limestone and Dolomite Deposition

Both the Upper Carbonate Member and the Hellnmaria Member are dolomite at the study sites. The Upper Carbonate Member is blue-gray dolomite, mostly burrowed, and the overlying Garden City Formation is a more orange or red laminated dolomite. The Hellnmaria Member is also a blue-gray dolomite that is heavily burrowed and is overlain by the Red Tops Member, which is an orange/red laminated dolomite with no visible trace fossils or stromatolites.

Limestone and dolomite provide important evidence that shallow marine seas covered much of the world during most of the Phanerozoic (Dickinson, 1992). Limestones are formed from precipitation of calcium carbonate minerals (calcite and aragonite) by calcareous plants and animals. This process is largely controlled by pH, which must be at least somewhat alkaline, and the partial pressure of dissolved carbon dioxide in the seawater. If equilibrium conditions are disturbed by loss of carbon dioxide, the concentration of hydrogen ions decreases and the pH increases, which is why ocean water is generally more alkaline than fresh water (Bathurst, 1975). The Late Cambrian was a time when an abundance of limestone was deposited in the Great Basin of Utah (Miller et al., 2011).

Calcite is precipitated in preference to aragonite at times when the concentration of magnesium ions in the ocean is low. These times of calcite precipitation have been linked to times of high rates of seafloor spreading, which increases the removal of magnesium from seawater by absorption into hot seafloor basalts (Stanley and Hardie, 1999).

Modern limestone deposition is occurring at a much slower rate than it has in the

Cambrian and the reason is partly due to dramatic climatic differences. There are two contrasting climatic states of Earth through geologic time—icehouse and greenhouse. These produce dramatic differences in seawater temperature and chemistry. During greenhouse conditions, there are no polar ice caps, sea level is higher, and the oceans are less stratified, and enormous volumes of limestone can be deposited. This apparently was the situation during the Late Cambrian in Utah, and it describes the temperature events and changes that were occurring during the time of deposition of the stromatolitic strata in this study (Poole, 1992). The Late Cambrian was a period of higher temperatures with no glaciation, because it occurred at the beginning of a greenhouse climate state, and this generally means a lack of polar ice and warmer seas (Poole, 1992; Whitaker and Xiao, 2010).

Lab experiments have shown that dolomite does not precipitate from aqueous solution at normal atmospheric conditions. This may partially explain why most of the dolomite found is Paleozoic in age, and possibly formed when the limestone was exposed to magnesium-rich fluids that replaced calcium ions (Bathurst, 1975; Whitaker and Xiao, 2010).

During the Late Cambrian, limestone was deposited in shallow marine water during a greenhouse state. Icehouse conditions exist today, so it is difficult to identify an exact modern limestone depositional environment that is analogous to that of the Late Cambrian in western North America.

Late Cambrian limestone was deposited in a shallow marine setting within the neritic zone (intertidal, inner shelf, and outer shelf) around the edges of a continent where siliciclastic sediment influx was low (Bernasconi, 1994).

Organisms play an important role in the formation of carbonate rocks, because they can extract dissolved components from seawater to manufacture shells and skeletons that are later incorporated into limestone and dolomite (Bathurst, 1975). Organisms also may play an important role in mineral precipitation, such as calcium carbonate, when their metabolism modifies the geochemical setting enough to cause mineral precipitation. (Whitaker and Xiao, 2010). An abundance of cyanobacteria, such as that which thrived in the Late Cambrian, possibly could create an environment favorable to the precipitation of calcium carbonate.

2.5 Methods

2.5.1 Fieldwork and Collection

Stratigraphic intervals of interest were studies along strike to document the lateral extent of formation contacts above and below the stromatolitic facies. These were noted and photographed along with exposed fossils and sedimentary structures. Specimens were studied and collected in both the South Lakeside Mountains and the southern House Range. The studied specimens were collected from *in situ* outcrops, and loose specimens lying on the ground as float material were avoided.

2.5.2 Thin Sections

Thin section study focused on the burrows and burrow margins in both stromatolites and dolomite. Burrow walls in thin section were evaluated using a polarizing microscope with transmitted light in order to analyze the petrographic differences between burrows, stromatolites, and dolomite or limestone matrix to enable a

better understanding of the sedimentary processes of the burrow system.

Each sample was cut, sanded, and polished using a diamond blade, and mounted in a 24 mm-round epoxy plug at the University of Utah Department of Geology and Geophysics Sample Preparation and Thin Section Laboratory. Samples were oriented vertically so that comparisons could be made between different samples in the same orientation. All the samples were photographed with a standard digital camera with 3x optical zoom. The thin sections were carbon coated at the University of Utah prior to QEMSCAN analysis.

2.5.3 QEMSCAN Analysis

Quantitative Evaluation of Minerals by Scanning Electron Microscopy (QEMSCAN) is an analytical tool used to acquire and obtain large amounts two-dimensional mineralogical data. This system is fully automated, allowing for statistically reliable and reproducible information to be obtained in a very short period of time. After analysis, the data assemblages can be used to help determine and understand mineral composition on a microscopic scale. It provides high-resolution data from polished thin section samples using automated point counting with resolution as low as 1 micron. Developed originally for the mining industry, QEMSCAN is based on a Carl Zeiss EVO 50 automated scanning electron microscope (SEM), with four EDS detectors that are used simultaneously (Haberlah et al., 2011). QEMSCAN matches characteristic x-ray spectra with those from a database of known or predefined elemental composition and mineral definitions defined in the form of a species identification profile (SIP).

QEMSCAN data were interpreted using the 710A primary SIP (species

identification protocol), modified from Barrick Gold Corporation's 7271 primary SIP, which is a general SIP that follows a set of standard mineral equations similar to XRD databases.

Once scanned, the data were uploaded into the iMeasure software, captured on a pixel scale, arranged in a grid pattern, and then individual grains are classified. Once this is done, each mineral is assigned a color code. Finally, iDiscover software produces a detailed false-color image created from 1 to 3 mm sized square fields that are analyzed, stitched together, and overlain onto the scanned area for easy reference to spatial location (Haberlah et al., 2011). Each color is a mineral assignment for each pixel. The analysis can be used to detect inclusions and cement phases present in a polished section of rock and to resolve boundary definitions and phases between mineral grains.

QEMSCAN analysis of the thin sections allowed details of the mineral chemistry and the sediment grain size, shape, and packing to be analyzed, thus shedding light on possible diagenetic aspects. QEMSCAN was useful for evaluating the texture and mineralogy of the sediment filling the burrow systems in order to determine if they were occupied as open burrows or were filled with sediment by the burrowers.

2.6 Ichnotaxonomy

The morphology of a trace is an expression of animal behavior and is also the basis of the name (Seilacher, 1967b).The same organism may produce more than one ichnotaxon, because organisms typically exhibit a multiplicity of behaviors. One example is when an organism is feeding, resting, and moving through sediments, all these behaviors can produce different ichnotaxa which means that one ichnotaxon may be

produced by several different organisms (Ekdale and Bromley, 1991). In many cases, trace fossil producers are soft-bodied organisms and not preserved. Trace fossils that record important evidence of activities of organisms may make up the bulk of the biomass in most benthic communities (Pemberton et al., 1992).

Trace fossils commonly are preserved in rock units that contain no body fossils, and because they are absolutely *in situ* in most cases, they often are useful as diagnostic facies fossils. There are rare examples of transported burrow fills, burrow lining, or borings in shells, but the reworking usually is detected easily due to mode of preservation (Seilacher, 2007). Because trace fossils tend to have long stratigraphic ranges and narrow environmental ranges, they are useful in many ways in the study of sedimentology and paleoecology.

The substrate also has a direct effect on how trace fossil structures are created and preserved. An organism creating a burrow in soft sediment may need to reinforce the burrow walls, but it may not need to do this in firmer (compacted) sediments, thus the same animal may create two different trace fossils in two different types of substrate (e.g., *Ophiomorpha* vs. *Thalassinoides*).

Although it is difficult to classify trace fossils purely on a morphological basis, Bromley (1967a, p. 166) states, "In the final analysis, it is the morphology of the trace as an expression of animal behavior that is the basis of the name". Ethology is an important consideration, but it generally is not used as a criterion when classifying trace fossils. The binominal scheme is used for ichnotaxonomy, but the terms ichnogenera and ichnospecies are used to differentiate between trace fossils and body fossils. The International Code of Zoological Nomenclature (ICZN) formally accepts ichnofamilies

(1999, Art. 10.3), although most ichnologists tend to avoid classifying particular trace fossils formally at the family (or higher) level.

2.7 Ethology of Trace Fossils

Although ethology of trace-making animals is not applied in ichnotaxonomy, it commonly is used to categorize trace fossils on the basis of the types of behavior that the trace fossils represent. Seilacher (1953) originally proposed five main ethologic categories: resting traces (cubichnia), locomotion traces (repichnia), grazing traces (pascichnia), feeding traces (fodinichnia), and dwelling traces (domichnia). Additional categories have been introduced subsequently, such as escape traces (fugichnia) added by Frey (1973), farming traces (agrichnia) added by Ekdale (1985), and trapping traces (irretichnia) added by Lehane and Ekdale (2014). There are even more categories recognized by some ichnologists today; the above list is just a partial list of commonly used terms. In this study, the trace fossil characteristics reflect dwelling traces (domichnia) and infaunal deposit-feeding traces (fodinichnia) (Figure 2.2).

2.8 *Thalassinoides* Morphology

The most commonly used characteristics to diagnose the ichnogenus *Thalassinoides* are: a branching network of tunnels with or without swelling at branching junctions, vertical shafts connecting to the surface of the sediment, regularity of branching forming "Y" and "T" shaped bifurcations, superimposed tunnels with a three-dimensional open frame, and circular to elliptical cross-sections of burrow tunnels (Ekdale and Bromley, 1991; Figure 2.3).

The *Thalassinoides* described in this study in the Late Cambrian of Utah represent an anastomosing 3-dimensional open framework architecture. Burrow width is similar throughout the entire burrow system, and the burrow fill is usually noticeably lighter in color than surrounding sediments (Figure 2.4). These burrows are branching, have a fill that is different in texture from surrounding sediments, and are geometrically irregular and constricted, so they are assigned to the ichnospecies *Thalassinoides paradoxicus*.

2.9 Quantitative Methods of Trace Fossil Identification

Quantitative methods in ichnology allow for objective comparisons between different trace fossils, and they aid in identification. Trace fossils are mostly grouped on the basis of their overall morphological appearance. To describe *Thalassinoides,* the burrow angle is a helpful ichnotaxonomic tool. Branching angle is morphology dependent (Lehane and Ekdale, 2014). Burrow diameter can also be useful, because it does not change significantly during diagenesis, except in the case of over-silicified burrows, or compaction. Burrow length is less useful, because the preserved length depends on many variables, including weathering and fracturing of samples.

Thalassinoides trace fossils can be described by their topology, which refers to the basic geometric configuration that remains invariant under bending and stretching (Lehane and Ekdale, 2014). This is referred to as the fundamental ground plan of the burrow structure. *Thalassinoides* consist of anastomosing network burrows, so they offer a large number of pathways for organism travel within the burrow system.

Previous workers have used quantitative methods to compare and contrast various types of trace fossils in an objective way (Droser and Bottjer, 1986, 1987; Magwood and

Ekdale, 1994). Quantitative approaches have proved helpful in analyzing bioturbation (Miller and Smail, 1997), vertebrate trackways (Bates et al., 2008; Thulborn, 1990), deep-sea cores (Ekdale et al., 1984), spatial patterns (Pemberton and Frey, 1984), feeding patterns (Lehane and Ekdale, 2013), and ichnotaxonomy (Orr, 1999; Uchman, 1995, 2003).

In a topological sense, trace fossils can be divided into three different forms: lines, trees, and networks (Lehane and Ekdale, 2014). *Thalassinoides* are network forms, which are similar to branching forms except that the branches are anastomosing (i.e., the branches split apart and then rejoin). The diagnostic morphology of *Thalassinoides* is most apparent when viewed parallel to the bedding plane, and when viewed this way, the "T" and "Y" tunnel junctions are much more apparent. The branching angle can reveal a great deal about the behavior of the organism that created the *Thalassinoides* burrow system. Branching angle may reveal the type of movement and/or the anatomical body plan of the burrower (Lehane and Ekdale, 2014).

Ichnofabric is defined as the sedimentary rock fabric caused by biogenic reworking (Ekdale et al., 1984). Ichnofabric index, *ii* (Figure 2.5), provides a useful tool for semiquantitative analysis of rock fabric that has been bioturbated by organisms, creating trace fossils prior to cementation of the rock (Droser and Bottjer, 1988). Most of the rock from both study areas has an *ii* index of between *ii*3 and *ii*5. Ichnofabric index 5 depicts completely bioturbated sediment. There is an ichnofabric index 6 in which the sediment is completely bioturbated and there are no burrow features or sedimentary features left at all (Droser and Bottjer, 1986).

2.10 Behavioral Convergence

Different trace makers may produce identical structures when behaving similarly. Trace fossils represent the actual path of the organism, and in most cases, the organisms are believed to fill most or all of the cross-sectional area of the burrow.

Thalassinoides generally are attributed to the burrowing activities of decapod crustaceans in most Mesozoic and Cenozoic occurrences. Early Paleozoic *Thalassinoides* are the exception because decapod crustaceans are unknown from that time. The first documented occurrence of decapod crustaceans is in the Devonian (Feldmann and Schweitzer, 2010). Paleozoic records of *Thalassinoides* are incomplete and rare, so it is unknown at this time what kinds of organisms created the Early Paleozoic *Thalassinoides*. Either the first record of decapods is incomplete, or some other type of organism must have been responsible for producing these Paleozoic *Thalassinoides,* and this would be an example of behavioral convergence (Carmona et al., 2004).

There was a large increase of infaunal trace fossils from the Precambrian to the Early Cambrian (Crimes, 1974). Extensive colonization of infaunal habitats occurred approximately 65 million years before the Cambrian-Ordovician boundary, and the intensity and diversity of bioturbation steadily increased during the rest of the Cambrian. It has been suggested that the maximum depth of bioturbation during the Late Cambrian was only about 6cm (Droser and Bottjer, 1988). The burrows observed in this study suggest greater depths of bioturbation.

2.11 Analytical Results

All the samples from both the Hellnmaria Member in the House Range and the Upper Carbonate Member in the South Lakeside Mountains have been recrystallized to dolomite or replaced with microcrystalline quartz. Because of the pervasive dolomitization, it is difficult to discern some details about the diagenesis that has affected the burrows. In order to examine the effects of diagenesis on the studied rocks, thin section microscopy and QEMSCAN analysis were conducted.

2.12 Thin Section Microscopy

All thin-sectioned specimens are from the Upper Cambrian strata containing a stromatolitic facies with *Thalassinoides* trace fossils. In the House Range, all specimens are from the Hellnmaria Member of the Notch Peak Formation, and those collected from the South Lakeside Mountains are from the Upper Carbonate Member of the St. Charles Formation.

Burrowed stromatolite samples were cut with a rock saw both parallel and perpendicular to the top of the bedding planes in order to better visualize the 3-dimensional box network structure. These cut slabs then were polished to bring out the color and textural qualities (Figure 2.6). All thin sections were made from the 10 samples that were serial sectioned.

2.13 Diagenesis of Silicified Burrows

Late Cambrian *Thalassinoides* vary greatly in size, depending on whether they are silicified or unsilicified. The silicified *Thalassinoides* are much larger than the un-

silicified counterparts, because the silicification process apparently spread beyond the burrow margin into matrix sediment. In this study, the smaller the silicified burrow structures are, the more closely they resemble the original morphology of the dolomitized burrows.

Chert begins to form during early diagenesis prior to complete compaction of the sediment (Schmidt, 1986). If dolomitization occurred first, then the burrow tunnels may have been too compacted and too low in porosity for the silica to be incorporated. In flints in Cretaceous chalk, the silica precipitated in sediments with porosities of approximately 75%-80% (Clayton, 1986). The *Thalassinoides* flint is diagenetic chert that apparently reflects a fabric contrast in porosity and permeability between the burrows and the surrounding sediments, which facilitated silicification of the burrows (Bromley and Ekdale, 1986).

2.14 Dolomitized *Thalassinoides*

Most of the burrows in the stromatolitic facies in both field areas are dolomite, and they are found inside dolomitized stromatolites (Figure 2.7). When viewed in outcrop, the dolomitized burrows are lighter in color than the surrounding sediments, but when viewed in thin section, the difference in fabric of the burrows and the surrounding sediments is much more evident. Dolomitization occurs during intermediate to deep burial partly because deep burial increased the temperature and raising the temperature reduces the amount of magnesium required for dolomitization (Bathurst, 1975; Whitaker and Xiao, 2010).

2.15 Microstylolites

       Thin section analysis shows that the *Thalassinoides* burrows have a definite burrow margin where the dolomite crystal size differs, and between these different crystal sizes appear to be micro-stylolites (Figure 2.8). Burrow margins all are located between the large and small dolomite crystals in each thin section. It is unclear why the crystals inside the burrows are larger, but this could be due to a difference in porosity and permeability of the reworked sediment inside the burrows

       Stylolites are pressure-solution features found in rock where individual grains have undergone dissolution where they come in contact with other grains. They are common in ancient carbonate rocks, such as the Cambrian of Utah. Grain-to-grain pressure is the driver that causes dissolution when the solubility product constant becomes high, due to the grain-to-grain contact under adequate pressure to create a concentration gradient (Bathurst, 1975). When the concentration gradient is established, the solute ions diffuse away from the vicinity of contact into less concentrated solution (Bathurst, 1975).

       Stylolitization changes the resultant thickness of rock units and the amplitude of the jagged edges indicates a minimum amount of vertical loss of thickness (Bathurst, 1975). Samples of stylolites in this study have amplitude of approximately 100 to 200 μm, so the loss of thickness is only significant when the total number of micro-stylolites is taken into consideration.

       Pressure-solution occurs when the directional pressure transmitted between the grains is greater than the hydrostatic pore pressure. The stylolites are formed when pressure-solution occurs after cementation of the rock and the dissolved ions are unable

to travel perpendicular to the stress azimuth. The ions form the jagged linear feature that stops when the solution runs out. They are the color of the particular ions present in the solution, and they commonly contain residual noncarbonate material rich in insoluble clays (Bathurst, 1975).

2.16 QEMSCAN Analyses

The QEMSCAN analyses verify that the silicified burrows are mostly quartz, but they contain small amounts of dolomite and calcite inside the quartz. In Figure (2.9), the quartz section is the burrow, and the light blue is dolomite surrounding the burrow. The burrow margin is shown to be dark blue, and this is evaluated as a mixture of various minerals, including iron and magnesium, creating the dark colored line shown in thin section as a stylolite.

The dolomitized stromatolites are only visible at the outcrop level, because weathering patterns and slight color difference bring out the structure of the stromatolite layers. This is also why they are must easier to see at the outcrop under low angle light conditions. When viewed in thin section or using QEMSCAN analysis, the stromatolite laminations are not visible due to overprinting from extensive dolomitization (Figure 2.10).

QEMSCAN verified that the unsilicified burrows found in all the rocks collected for this study are almost completely dolomite. This is a case of dolomitized burrows not found in fossilized stromatolites. The difference between the burrows and the sediment is the size of the dolomite crystals. The larger crystals are inside the burrows, and the

smaller crystals are in the rock surrounding the burrows. Because everything is dolomite, QEMSCAN analysis shows a fairly homogeneous geochemical pattern (Figure 2.11).

2.17 *Matherella* and *Matthevia*

In the southern House Range, many silicified body fossils of polyplacophoran mollusks (*Matthevia walcotti*) and gastropods (*Matherella wahwahensis*) are found on tops of the burrowed stromatolites, concentrated along the bedding planes (Figure 2.12). These body fossils are not found inside the burrow tunnels, so there is no reason to conjecture that these animals were the burrowers. They probably were epifaunal grazers feeding on the surfaces of the stromatolites. These fossil taxa actually are quite rare in most Late Cambrian settings (Vendrasco and Runnegar, 2004), but they are found in abundance in this location, and they make up virtually all of the macrofossils in this stromatolite facies.

All *Matthevia* and *Matherella* specimens observed and collected in the southern House Range from the Hellnmaria Member have been preserved by silicification; the original aragonite shells have been replaced by quartz. It is interesting that these fossils and many of the burrow tunnels became silicified, and this could be a clue into some of the diagenetic aspects of this field area.

2.18 Discussion

Stromatolites make up much of the rock containing the burrow systems. They are formed in shallow water by microbial biofilms, probably cyanobacteria, by trapping and cementing grains of rock into biostratification structures. Stromatolites are among the

oldest fossils on Earth, having a range from approximately 3.5 million years ago to modern day (Shapiro, 2007). There is some debate about whether stromatolites should be referred to as biogenic sedimentary structures or physical sedimentary structures. They are results of a living organism (cyanobacteria), but many workers debate whether cyanobacteria can actually exhibit behavior (Shapiro, 2007). The organic-rich lime mud in and around stromatolites presumably offered an abundant food source for the deposit-feeding animals that created the *Thalassinoides* burrows. A highly alkaline setting would allow the stromatolites to grow without being destroyed by an abundance of grazers.

Thalassinoides paradoxicus in this study is both a dwelling trace (domichnion) and deposit feeding trace (fodinichnion) constructed by infaunal organisms. The probable burrowers most likely were mining the shallow marine mud for food as well as seeking protection in the complex web of burrows. The burrows in this study appear to have been passively filled, open burrows, revealed by contrasting color with host sediments. The trace fossil characteristics reflect dwelling (domichnia) and infaunal deposit-feeding (fodinichnia) behavior. The feeding strategies of the organisms in this study are mostly associated with stromatolites, and the resulting trace fossils are always autochthonous, meaning that there is no evidence of transport of the burrows. This evidence suggests that these open-framework burrow systems may be a record of the behavior of some of the world's first infaunal deep deposit feeders.

As a result of dolomitization and deformation due to age and burial of the strata, many diagnostic features are absent from the Upper Cambrian *Thalassinoides*. Missing are uniformity of tunnel diameters, regularity of branching angles, and bioglyphs (wall scratchings). The bioglyphs are usually apparent in undeformed specimens, because

post-Paleozoic *Thalassinoides* is generally considered to be the dwelling structure of decapod crustaceans, including the Superfamily Thalassinidae (Carmona et al., 2004).

Decapods crustaceans in the Cenozoic and Mesozoic made *Thalassinoides paradoxicus* branching burrow systems, similar to these in the Cambrian, but body fossils of decapods do not appear in the fossil record until the Devonian. Worms and lingulid brachiopods do not make highly branched or anastomosed burrow systems, and trilobites did not have a suitable anatomy to make highly branched or anastomosing burrow systems. Other common Late Cambrian animals known to have existed then, such as archeogastropod and polyplacophoran mollusks, were obvious nonburrowers. Although conodonts are abundant in bulk samples of these Upper Cambrian carbonate rocks (Miller, 2011), there is no direct evidence to suggest that conodont animals were burrowers and possible creators of the *Thalassinoides* burrows. Thus, at this point, it must be said that the biological identity of the *Thalassinoides* burrowers in these stromatolitic rocks remains unknown.

The burrows are passively filled, and the different color of the burrows from the surrounding sediment suggests that they were open burrows that were filled with altered sediments. Thin section and QEMSCAN analyses verify that the sediment fill in the *Thalassinoides* burrows is different than the surrounding sediment. The substrate had an effect on how these trace fossils were created and preserved, and the burrow architect must have lived in relatively firm sediment, since the burrow margins show no signs of reinforcement.

In this study, *Thalassinoides* are not readily apparent when viewed perpendicular to the bedding plane. The bifurcations are "Y"- or "T"-shaped. The burrow walls are

smooth, but pervasive dolomitization may have obscured fine structures. Orientation and spacing between burrows is not random; the burrows commonly are evenly spaced both vertically and horizontally. The burrows systems are obvious in outcrop, because they usually are revealed by contrasting color. When viewed in thin section, the burrow fill typically has a different texture and fabric than the surrounding sediment (Figure 2.13). According to Sheehan and Schiefelbein (1984), *Thalassinoide*s burrows are rarely compressed, and when they are compressed, the system tends to simply collapse. The absence of compression and occasional collapsed burrows are further evidence that the *Thalassinoide*s burrows were filled with sediment after being vacated by the occupant (Sheehan and Schiefelbein, 1984). The sediment filling of the burrows is evidence suggesting that this was the work of an infaunal deposit feeder.

Stylolites probably formed as a result of the differences in textures between the burrows and host sediment. QEMSCAN analysis shows the presence of iron and magnesium in the burrow margins, and this is likely the source of the dark color in the stylolites present here. The extensive micro-stylolites are evidence of overburden and burial of both Upper Cambrian units in this study.

2.19 Burrowed Cores of Dolomitized Stromatolites

In both field areas, the burrow fill, fossilized stromatolites, and surrounding sediment all have been dolomitized. Even though they are all the same type of rock, the sediment inside the burrow tunnels usually is noticeably lighter in color than the surrounding sediment in which the burrows are found. In the case of the over-silicified burrows, the burrows are a dark charcoal gray to black on the inside, and they contain a

weathered rind in various hues of brown and orange. These over-silicified burrows stand out three-dimensionally in junction preservation against the lighter-colored bedding plane due to differential weathering (Figure 2.14).

Dolomitization is a complex process in which magnesium ions eventually replace some calcium ions in calcite and aragonite (Bathhurst, 1975; Vasconcelos et al., 1995). This process involves a substantial amount of recrystallization causing diagenetic overprinting to be common. Petrographic studies in this study demonstrate that the rock in both field areas has been dolomitized extensively, and the original fabric of the rock has been overprinted. This is a result of diagenetic replacement by dolomite, not primary production of dolomite. This is one reason why the stromatolites can be observed at the outcrop scale, but they are not evident in the serial slabs or in the thin sections. The stromatolites have maintained some visible structure on a large scale, but in thin sections all that can be observed is the limestone that has been recrystallized into dolomite. Aragonite is a more metastable form of calcium carbonate (Bathurst 1975; Whitaker, 2010), and the aragonitic shells of the mollusks were preferentially recrystallized in the samples collected in this study.

2.20 Environmental Implications of *Thalassinoides*

The *Thalassinoides* in this study are far too old to have been created by decapod crustaceans (Feldmann and Schweitzer, 2010), so this eliminates the possibility of inferring their environment using knowledge of the habitat requirements of a known organism. Because trace fossils are almost always autochthonous, it can be assumed that they are found in the strata in which they formed. Seilacher (1967b) suggested that

recurrent assemblages of trace fossil taxa have paleoenvironmental implications, which led him to establish to the concept of ichnofacies, which is a characteristic association of trace fossils directly reflecting certain environmental conditions. *Thalassinoides* is a facies-breaking ichnogenus, meaning that it can be found in a broad spectrum of paleoenvironments. However, because the *Thalassinoides* found in this study are associated with cyanobacterial stromatolites, we have definite evidence of their specific depositional environment.

Bathymetry, salinity, and substrate are examples of the environmental conditions, which may influence the distribution of ichnofacies (Ekdale, 1988). The *Thalassinoides* in this study are unlined and not reinforced in any way, so they would have required a firmground substrate or else the burrows would collapse immediately. *Thalassinoides* are often included within the *Glossifungites* Ichnofacies, which is found in marine firmgrounds, which are typically stiff, compacted muds, into which suspension-feeders excavate their burrows (Pemberton and Frey, 1985). Ichnologists realize that distribution of trace fossils may be directly related to water depth, but this is not always the case with *Thalassinoides* (Ekdale and Bromley, 1984). Trace fossils are distributed according to bathymetry only as other ecologic factors controlling the distribution of trace-making organisms are correlated with bathymetry (Ekdale, 1988). Organisms that create trace fossils in marine environments typically are distributed according to three main factors: water depth, water chemistry, and substrate character. In marine settings, as the water becomes deeper, there is a decrease in temperature, light, wave energy, sediment rate, sediment grain size, and food supply (Ekdale, 1988).

Thalassinoides can occur in association with the *Skolithos* Ichnofacies or the

Glossifungites Ichnofacies, depending on whether the substrate was soft or firm,

regardless of water depth. The *Thalassinoides* in the Upper Cambrian of the Great Basin

are associated with firmground substrates in lime muds. Because the *Thalassinoides* in

this study are associated with stromatolites, it is evident that the paleoenvironment was a

shallow-marine, restricted platform. Stromatolites will break apart if they are subjected to

high-energy conditions, and they need clear, shallow water in order for the cyanobacteria

to perform photosynthesis. In a high-energy, high-oxygen, shallow-marine environment,

such as a beach, *Thalassinoides*-making organisms would find it difficult to excavate

unlined burrows, because their burrow systems would continually collapse due to the

constantly shifting substrate. This might be the main reason why the majority of the

Thalassinoides burrows documented in this study from the Great Basin are found inside

or very near stromatolites, and also why the stromatolitic facies in this study are more

heavily burrowed than the surrounding rock.

2.21 Over-silicified Burrows

Thin sections of the silicified burrows show microcrystalline quartz with a few

dolomite rhomboids incorporated within the quartz (Figure 2.15). This type of burrow

preservation presents itself as a nodular-shaped, three-dimensional burrow parallel to the

bedding plane. The burrows were approximately 2 cm in diameter before they became

silicified, but they average approximately 10 cm in diameter after being silicified. The

silicified burrows in the Southern Lakeside Mountains are lighter in color than the

silicified burrows in the southern House Range, probably due at least in part to

differences in weathering (Figure 2.16). The burrows that are over-silicified have lost

some of their original architecture, but it is evident that the silicified burrows are inside the rock as well as on the bedding plane (Figure 2.17).

The silica appears to be early diagenetic in origin. Although the initiual source of the silica is unknown, the silica may have come from opaline sponge spicules. As already mentioned, the majority of the microcrystalline chert nodules extend beyond the boundaries of the original trace fossil. It is believed that this is due to the silica growth continuing around the original burrow surface to form a nucleation site until the system runs low on silica (Bromley and Ekdale, 1984).

Thin sections were examined using a polarizing microscope with transmitted light to analyze the petrographic differences to better understand the sedimentary processes of the burrow system. Thin section analysis of burrows in dolomitized stromatolites revealed the relationship between the burrow margins and host sediment. Thin sections of samples of over-silicified burrows in dolomitized stromatolites also were examined.

2.22 Dolomite Crystals in Silicified Burrows

In many cases, the silicified burrow tunnels have dolomite crystals incorporated within the silica. During the silicification of Upper Cretaceous chalks in Europe, it was determined that they originally were composed of opal-CT, a soft, deformable solid that was not firmly cemented until precipitation of an intermediate opal-CT/chalcedony stage (Schmidt, 1986). In the Cambrian rocks, while in a semisolid state, dolomite was precipitated inside the silica structure, thus explaining the occurrence of euhedral dolomite crystals inside the chert nodules that represent the over-silicified burrows. The interior of the over-silicified *Thalassinoides* burrow is quartz with some dolomite crystals

visible in thin sections (Figure 2.15). The precipitation of the dolomite may have occurred during the phase when the silica was in a soft, deformable state. This is evidence that the burrow transformation to silica occurred later in the burial stage after the rock had undergone at least partial dolomitization.

Stylolites are a common feature in the stromatolitic facies in both the Hellnmaria and Upper Carbonate Members. They are mostly visible as micro-stylolites in thin sections. The stylolites formed between the burrows and the host sediment (in the burrow margins), because there is a difference in the porosity and permeability between the sediments. The thin sections show large crystals in the burrows and smaller crystals surrounding the burrow, which may be due to how diagenesis reacted to burrow sediment versus host sediments.

The stylolites in both field areas are mostly horizontal, thus inferring stress and pressure from overburden. Dolomite is highly resistant to pressure-solution (Mattes and Mountjoy, 1980), so the stylolites in these two field areas were probably formed prior to extensive dolomitization.

This is evidence that the burrow fill sediment was different in texture than the surrounding sediment. This may be evidence of the burrow being open and filled with different sediment that outside of the burrows, and then when the pressure-solution formed because of overburden, the stylolites formed a line delineating the burrow walls The ions that form the jagged linear feature are the color of the particular ions present in the solution, and this material commonly contains residual clay.

2.23 Conclusions

This study demonstrates that the structures located within and between individual stromatolites, observed and collected from both the Hellnmaria Member and the Upper Carbonate Member in the Great Basin, are burrows. They have an anastomosing, 3-dimensional, open framework construction. The burrow tunnel widths are more or less uniform within each system, and the burrow margins are observable. There are "T" and "Y" junctions in the burrow networks, and the burrows are revealed by contrasts in color and mineralogy (Figure 2.18). Because the burrow tunnels are geometrically irregular and constricted, they are assigned to *Thalassinoides paradoxicus*.

This limestone in both the Hellnmaria Member and the St. Charles Formation has undergone extensive dolomitization. This is not unusual, because a vast majority of Cambrian limestone is dolomitized. The burrow fill sediment and host sediment are not the same texture of dolomite. The burrowers were unknown infaunal, sediment-ingesting deposit feeders, as opposed to filter feeders or surface herbivores. Deposit feeders typically create open burrows that become filled with sediment piped down from above and possibly reworked by the organisms, thus changing the sediment properties. The *Thalassinoides* in this study are approximately 500 million years old according to the strata they are found in, so they are more than 100 million years older than the first fossil records of decapods, which are the presumed trace makers of *Thalassinoides* in younger rocks. Behavioral convergence of unrelated trace makers is the best explanation for the existence of these Upper Cambrian *Thalassinoides* from both field areas.

Dolomite crystals inside the burrows tend to be larger than those in the host sediment, possibly related to a higher porosity and permeability of the original burrow fill

sediment than that of the surrounding mud. Cambrian limestones often are totally dolomitized, apparently occurring simultaneously (at least in part) with preferential silicification of mollusk body fossils and burrows.

Many of the *Thalassinoides* burrows have been replaced with chert, which appears to be early diagenetic in origin. Most burrows are over-silicified due to the fact that the majority of the microcrystalline chert nodules extend beyond the boundaries of the original trace fossil. This is probably due to the silica growth continuing outward from the original burrow surface from a nucleation site until the system runs low on silica.
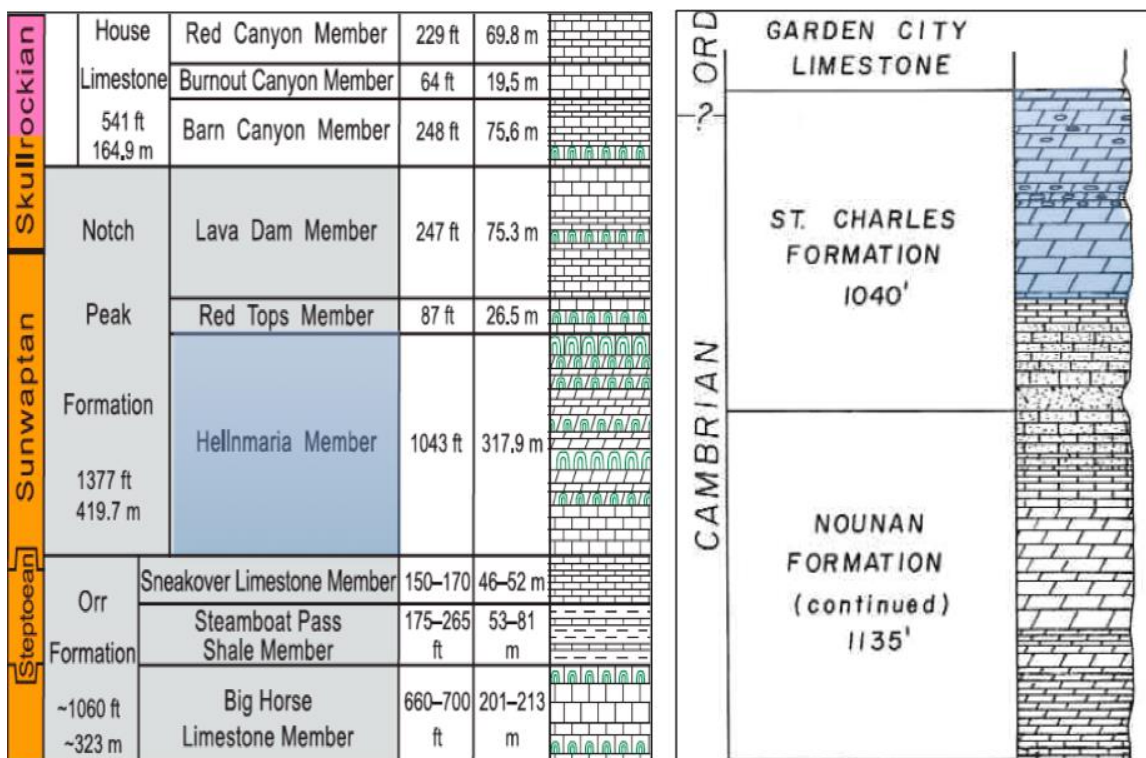
Figure 2.1. The stratigraphic column on the left illustrates the strata of the southern House Range with the light blue shaded area denoting the study area. The orange shaded area is Cambrian rock, and the pink shaded section shows Ordovician rock (modified after Miller, 2012). The stratigraphic column on the right shows the rock layers in the Upper Carbonate Member in the South Lakeside Mountains shaded in blue; this is where rock samples were observed and collected from this field area (modified after Stokes, 1986).
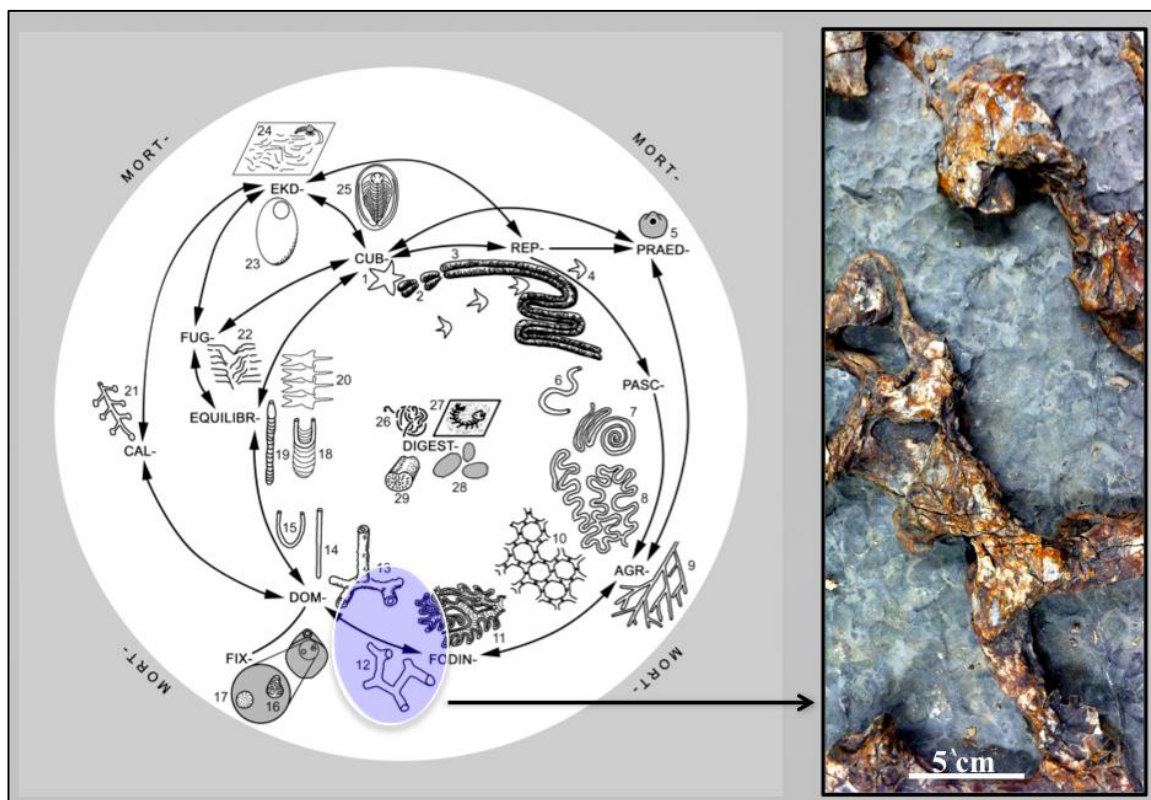
Figure 2.2. Circle diagram of ethologic categories of trace fossils showing an idealized drawing of a *Thalassinoides* burrow as an example of a fodinichnia trace fossil (12) on the left compared to the *Thalassinoides* burrows found at the South Lakeside Mountains field site (modified after Bromley, 1996).
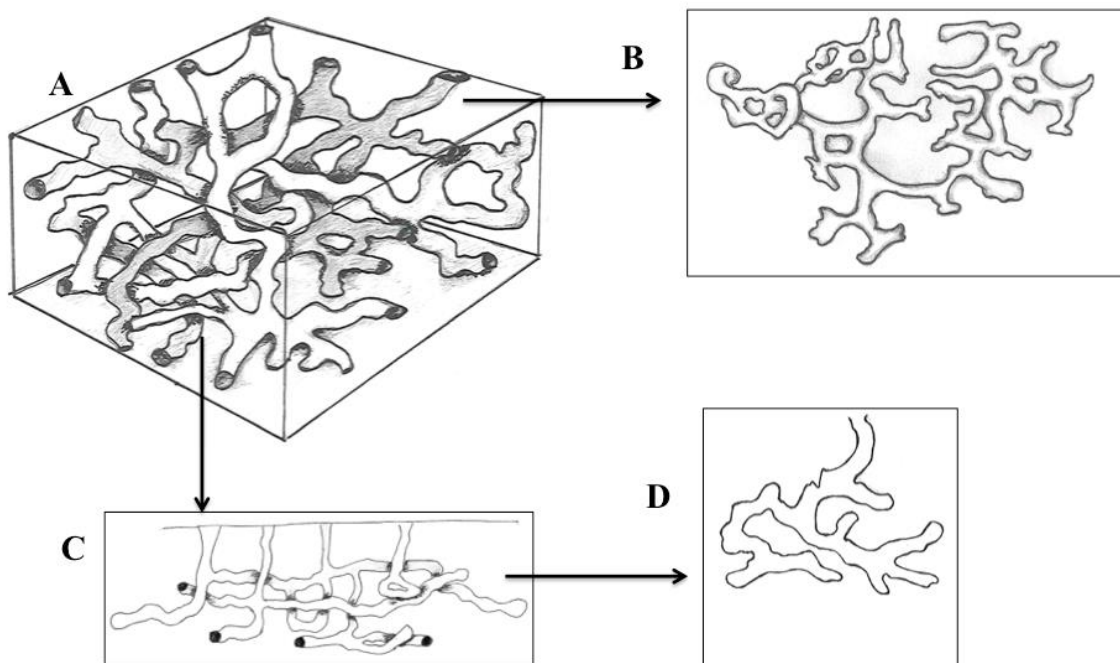
Figure 2.3 Series of drawings showing various views and dimensions of a typical *Thalassinoides* burrow system.

(A) Idealized 3-dimensional drawing of the box network burrow system (modified from Ekdale, 1984). Burrows can be silicified or unsilicified, and both varieties have the same basic morphology, but the burrow diameter size is much different.

(B) Drawing showing an idealized pattern of *Thalassinoides paradoxicus* viewed in 2 dimensions, parallel to the bedding plane. The anastomosing pattern is shown and depicts the organisms taking various pathways throughout the box network burrow system. This is a drawing of over-silicified burrows collected in the South Lakeside Mountains.

(C) Idealized drawing of view of the 3-dimensional boxwork burrow when viewed perpendicular to bedding (side view). This is a 3-dimensional view that shows some depth into the burrow structure.

(D) Drawing showing how the side view appears when viewed at the outcrop or in collected samples.

Figure 2.4. Burrowed stromatolite observed in vertical view in an outcrop in the southern House Range at the Steamboat Pass location from the Hellnmaria Member. The round lighter colored material represents the burrow opening and the elongated light colored sections are the side view of burrows. Both are visible in this dolomite because of the box network style of burrow.
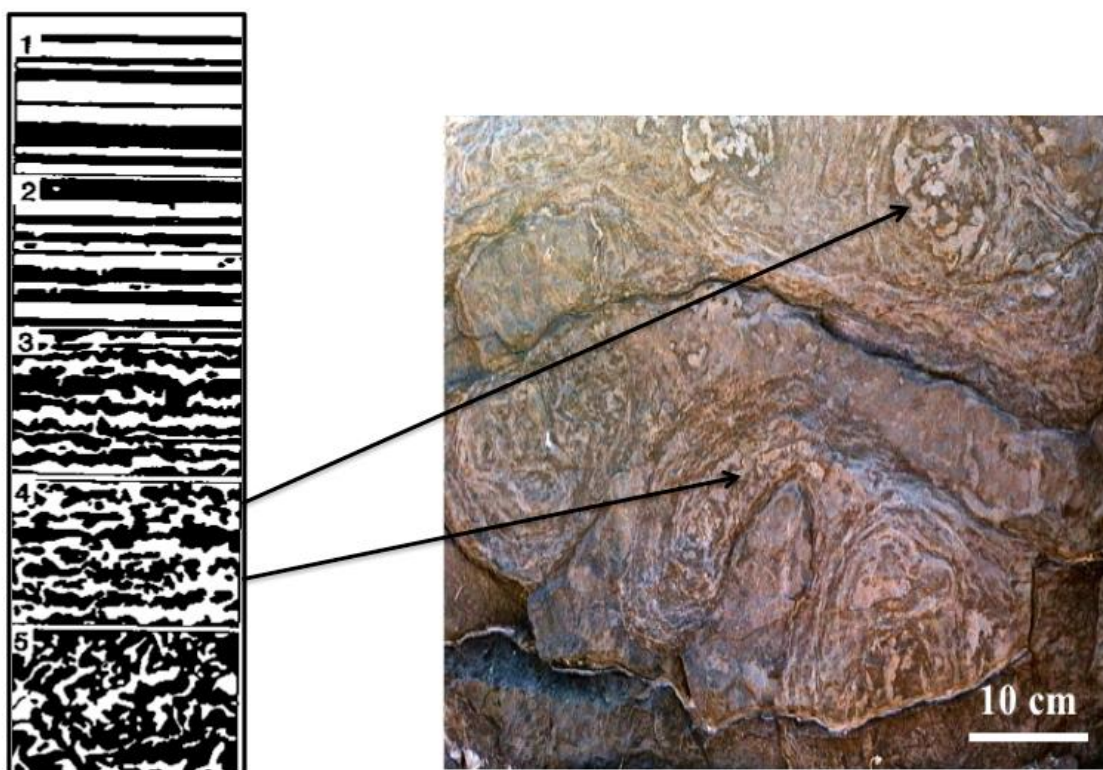
Figure 2.5. The diagram on the left is the guide for ichnofabric index for *Thalassinoides* (modified after Droser and Bottjer, 1988). The photo on the right was taken at the South Lakeside Mountains site, and shows an ichnofabric index of about *ii*4. The burrow tunnels are lighter colored than the surrounding dolomite.

Figure 2.6. Polished slab (cut vertically) of burrowed, dolomitized stromatolite from the southern House Range, collected at Steamboat Pass from the Hellnmaria Member. This photo shows the vertical and horizontal aspects of an unsilicified burrow system, and the sub-round cross-sectional shape of the burrow tunnels is evident. The unsilicified burrows are difficult to see even in this polished sample.
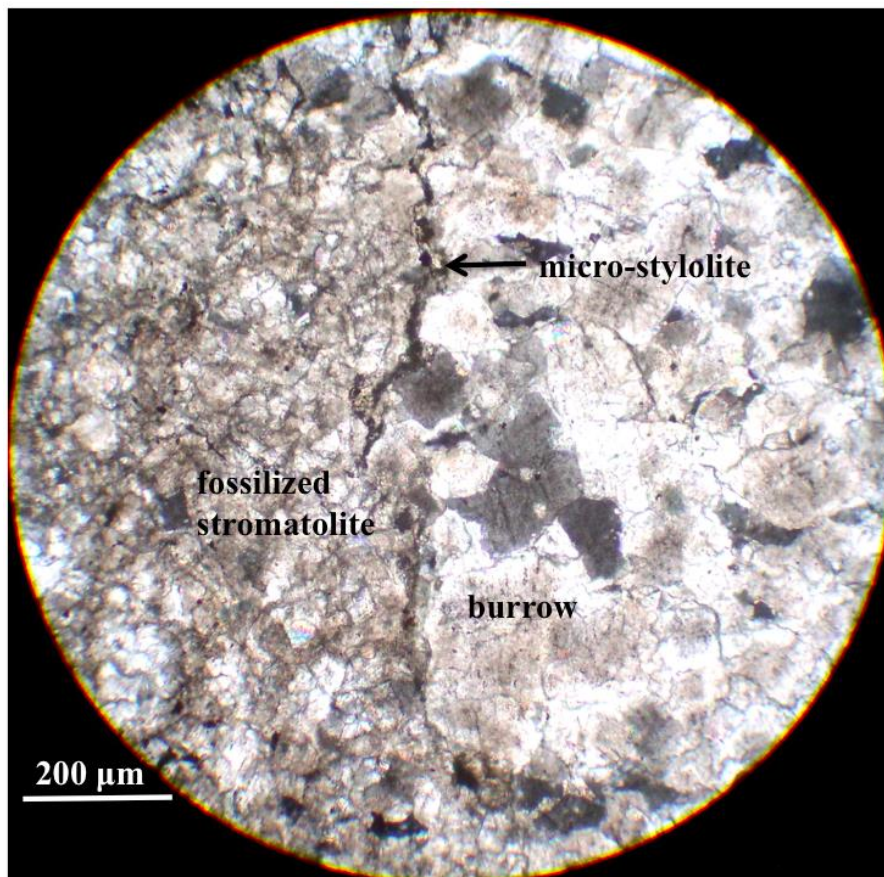
Figure 2.7. Thin section viewed under polarized light. The larger anhedral and euhedral dolomite crystals are inside the burrow. The host sediment is seen as smaller crystals. This is a thin section of a burrow in a dolomitized stromatolite. There is a stylolite (black arrow) between the two sizes of dolomite crystals.
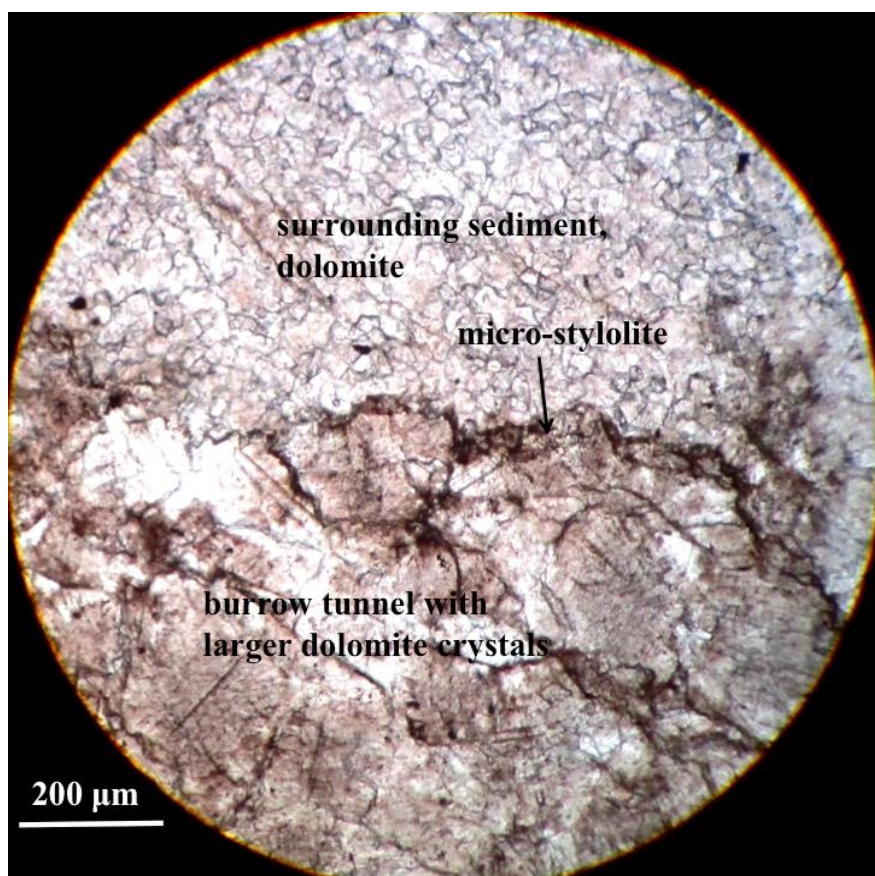
Figure 2.8. This section shlowing micro-stylolites located between dolomitized sections of the sample with the larger crystals in the burrows and the smaller crystals in the surrounding matrix. Micro-stylolite has the characteristic jagged dark line that defines stylolites, due to the color differnce created by the pressure-solution of the minerals iron and magnesium. Planar light view of thin section with a burrow in dolomite, which is not stromatolitic. The micro-stylolites are visible between the small and large dolomite crystals in the burrow margin.

Mineral Name	Mineral %
Quartz	51.75
Dolomite	47.05
Amphibole	0.63
Calcite	0.20
Apatite	0.12
Other_Silicates	0.11
Nontronite(LOW)	0.04
Micas	0.04
Olivine	0.02
Fe-Ox/CO3	0.01
Epidote	0.01
Background	0.01
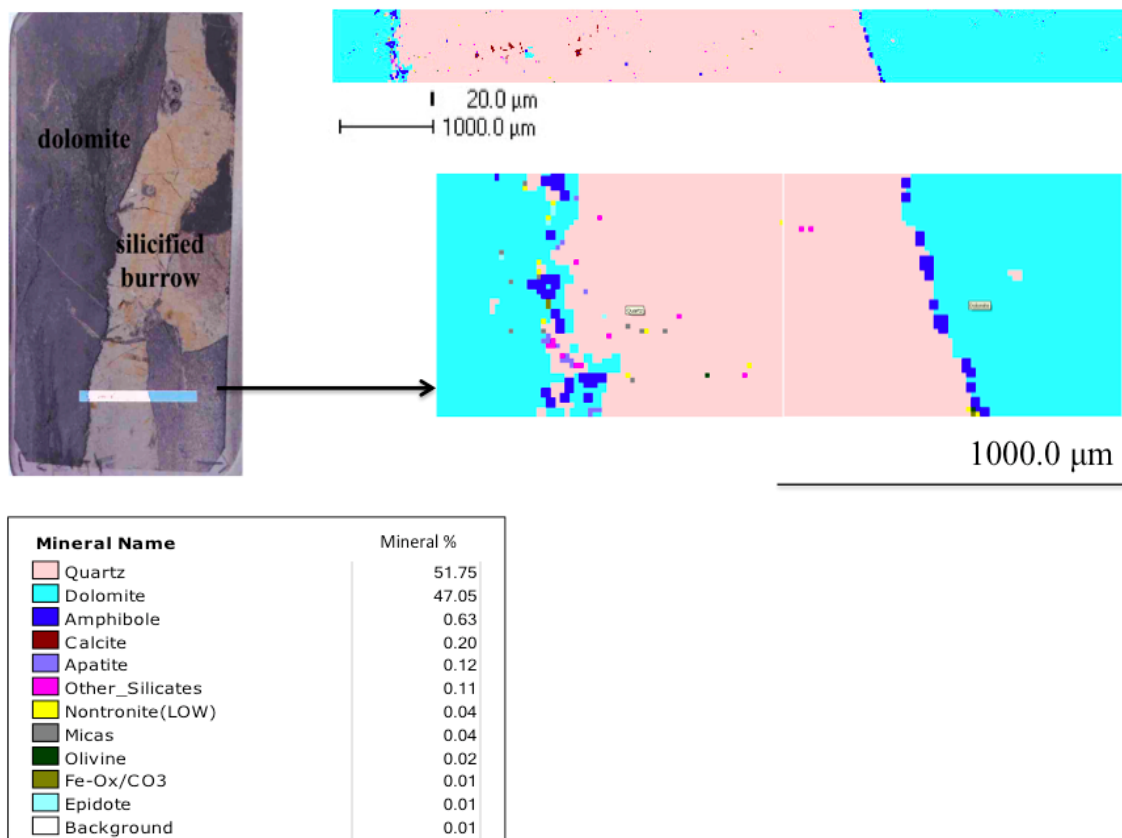
Figure 2.9. QEMSCAN image and mineral percentages of part of a burrow. The light colored bars that cross the burrow margins in the thin section photo show the section of the thin section that was further analyzed with QEMCSAN analysis. QEMSCAN results are the bars and chart shown above. The pink sections are assigned to quartz, and the light blue is assigned to dolomite. There are some dark red specks inside the quartz region, which results show are calcite, and there are a few dolomite specks scattered amongst the quartz. The darkest blue is where the stylolites are seen in thin section. The mineral chart reveals the percentage of each mineral found in the total bar analyzed

Mineral Name	Mineral %
Dolomite	98.80
Amphibole	1.02
Background	0.11
Apatite	0.05
Calcite	0.04
Quartz	0.04
Plagioclase	0.02
Sulphates	0.01
Fe-Ox/CO3	0.01
Mg(Fe,Mn)silicate	0.00
Goethite	0.00
Micas	0.00
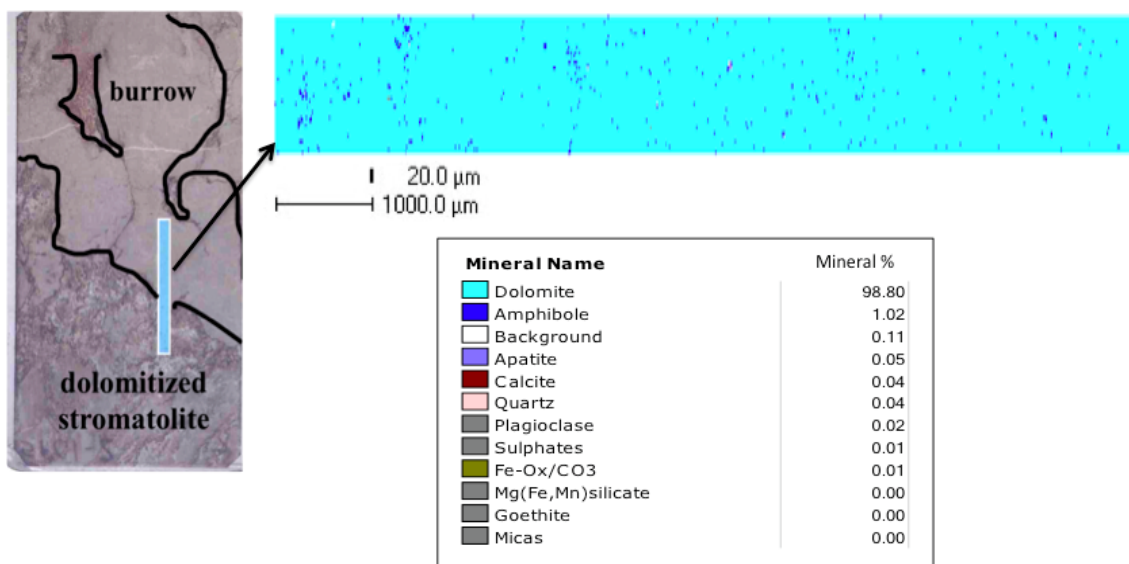
Figure 2.10. QEMSCAN image and mineral percentages of part of a burrow. The photo of the thin section above was cut from a polished slab of burrowed, dolomitized stromatolite. The blue bar on the photo is the section analyzed with QEMCSANS. The light colored bars that cross the burrow margins in the thin section photo show the section that was further analyzed with QEMCSAN analysis.   The QEMSCAN results reveal that the entire rock, including the burrow tunnels, has been dolomitized. The chart in this figure shows mineral percentages determined by the QEMSCAN analysis.
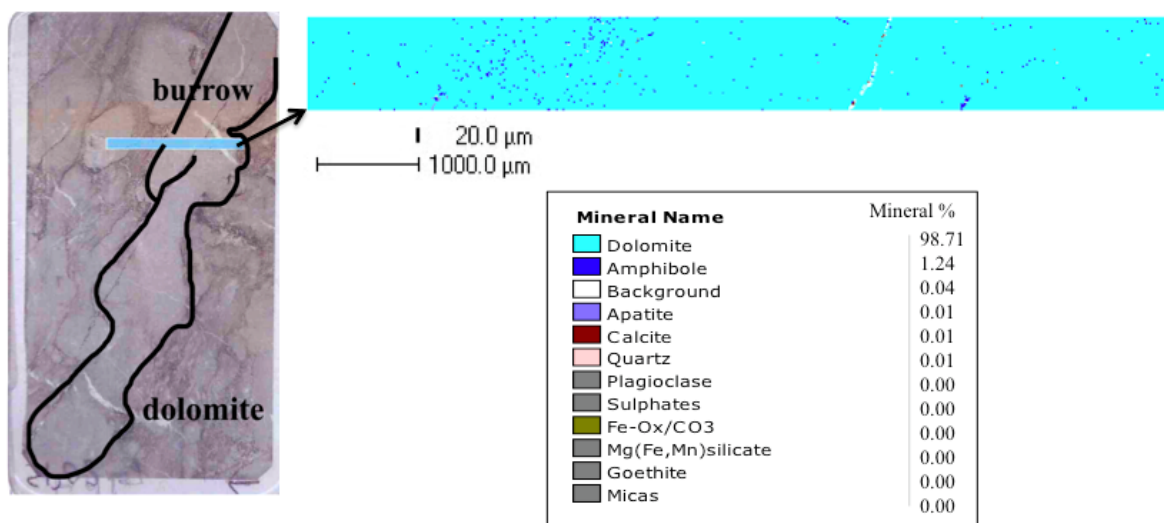
Figure 2.11. QEMSCAN image and mineral percentages of part of a burrow. The photo is a thin section of a burrow in dolomite, not a burrow in a fossilized stromatolite. The light blue bar in the photo denotes the section analyzed by QEMSCAN. The light blue is assigned to dolomite and the darker blue specks are a variety of minerals including iron and magnesium that are labeled as amphibole by the QEMSCAN analysis. The dark specks are in sections that are near or at burrow margins. The chart denotes the percentage of each mineral in the section as revealed by the QEMSCAN analysis.

Matthevia walcotti

Matherella wahwahensis
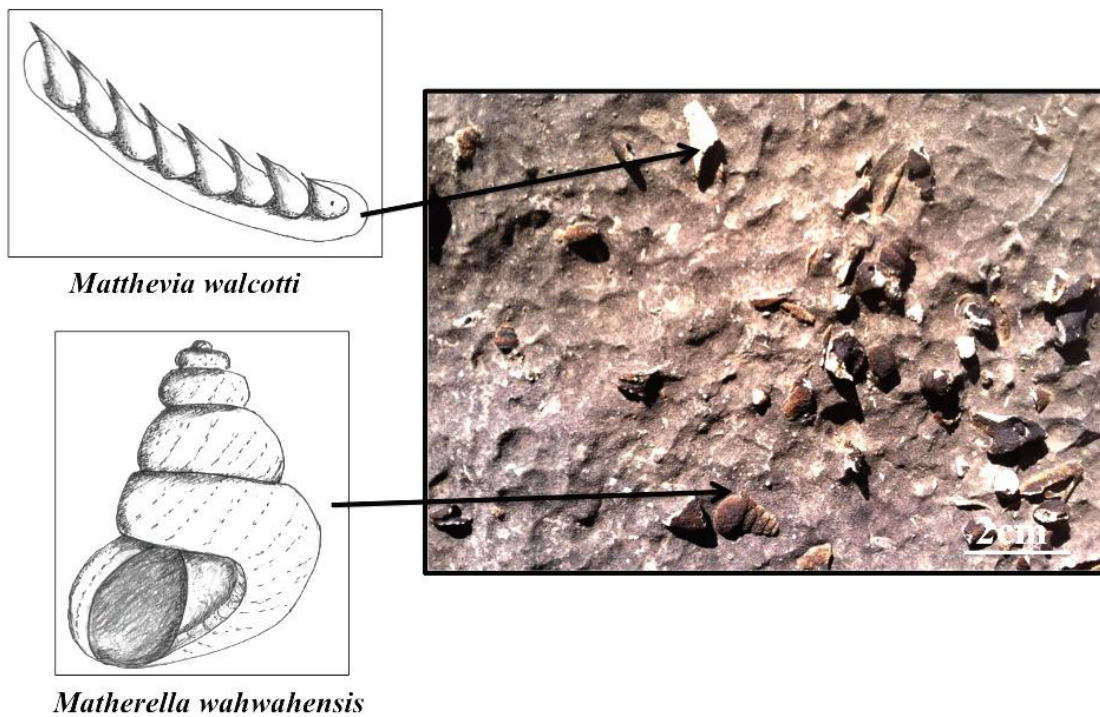
Figure 2.12. Silicified body fossils of polyplacophorans (Matthevia) & gastropods (Matherella) in the Upper Cambrian, Notch Peak Formation, House Range, Utah. These body fossils are located on top of the burrowed stromatolites in the field area. On the left, the upper drawing is Matthevia walcotti, a polyplacophoran mollusk (chiton), which is preserved by silicification in both field areas. Also on the left, the lower drawing is the sinistral gastropod, Mattherella wahwahensis, preserved by silicification in sediments of the Late Cambrian Hellnmaria Member and Upper Carbonate Member in western Utah.
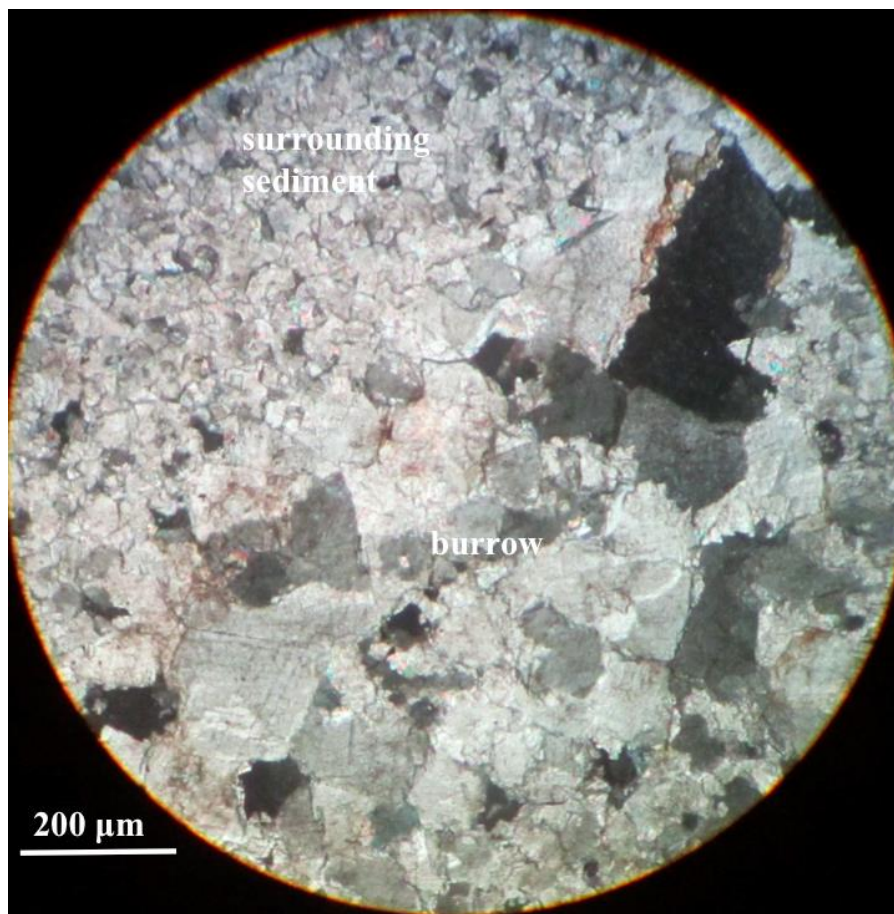
Figure 2.13. Thin section of a burrow and the surrounding sediment viewed under polarized light showing textural and fabric differences. The burrow tunnel has much larger dolomite crystals than does the surrounding sediment.

Figure 2.14. Photo of a bedding plane of the Hellnmaria Member at Steamboat Pass in the southern House Range of burrowed stromatolites. The dolomite burrows are lighter colored than the surrounding sediments, and the silicified burrows are more colorful and stand out 3-dimensionally from the surrounding sediments.
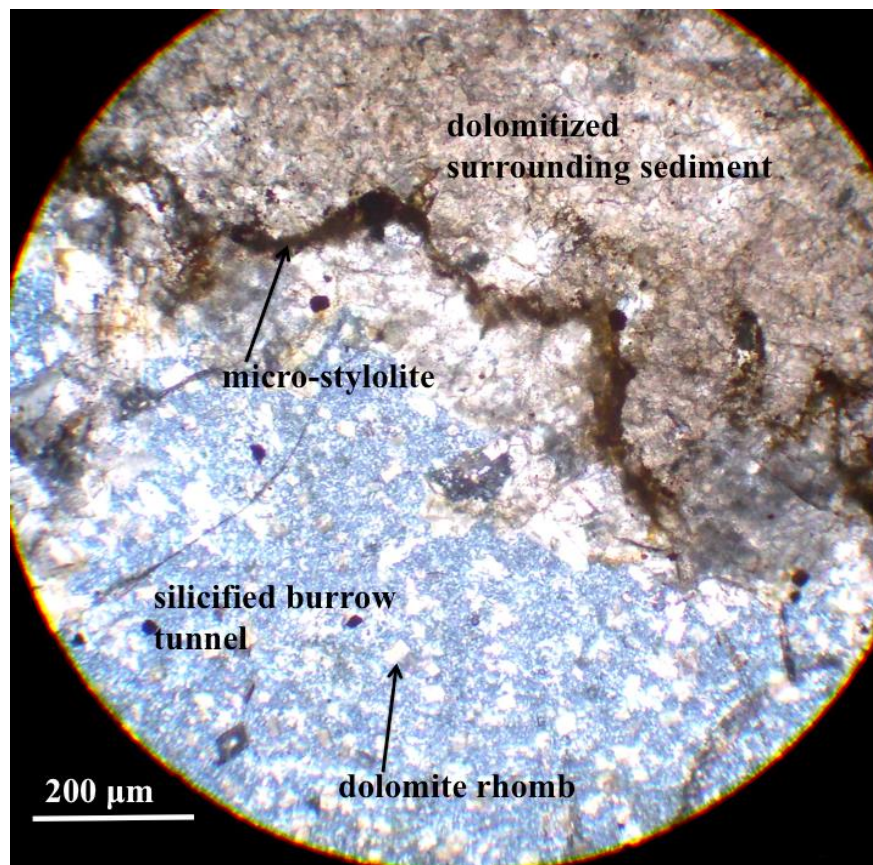
Figure 2.15. Thin section of an over-silicified *Thalassinoides* showing calcite and dolomite crystals inside the recrystallized quartz burrow. Burrow margins appear to be a micro-stylolite. The burrow and the sediment shown under polarized light demostrates the petrographic differences between the crystals sizes, shapes, and packing of the dolomitized features.

Figure 2.16 Photos showing contrasting colors of burrows and sediments in both field areas. (A) Over-silicified *Thalassinoides* burrows on a bedding plane in the St. Charles Formation, South Lakeside Mountains. (B) Over-silicified *Thalassinoides* burrows on a bedding plane in the southern House Range.
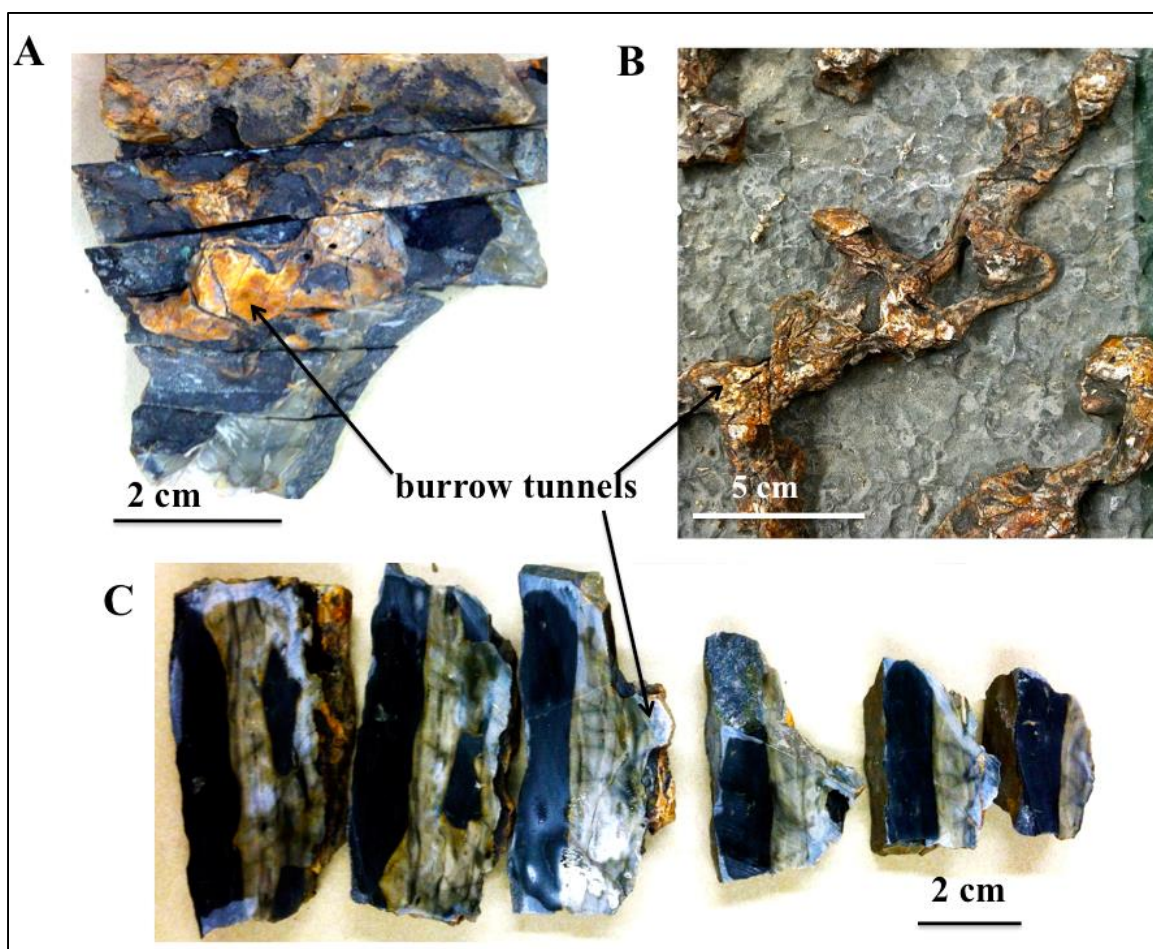
Figure 2.17.Series of photos showing how the silicified burrows appear in the rock and
how they appear once they are separated from the rock and serial sliced. (A) A silicified
Thalassinoides burrow sliced into 6 serial pieces. (B) Photo of a *Thalassinoides* burrow
system in the dolomite rock before being serial sliced. (C) Burrow tunnels after they have
been sliced and laid out in order. The dark colored section is the dolomite, and the lighter
colored rock is the silicified burrow tunnel.

Figure 2.18. Over-silicified *Thalassinoides* burrows on a horizontal bed surface in the southern House Range. They chert is black on the outside where it has weathered and light brown inside the weathering rind. QEMSCAN analysis shows that the limestone has been dolomitized. Hellnmaria Member, southern House Range.

2.24 References

BATES, K. T., P. L. MANNING, B. VILA AND D. HODGETTS. (2008). Three-dimensional modeling and analysis of dinosaur trackways. Palaeontology, 51:999–1010.

BATHHURST, R. G. C. 1975. Carbonate Sediments and Their Diagenesis (second edition). Developments in Sedimentology 12. Amsterdam, Elsevier, 459–473 p.

BROMLEY, R. G. 1967. Some observations on burrows of *Thalassinidean Crustaea* in chalk hardgrounds. Geological Society of London Quarterly Journal, 123:157–182.

BROMLEY, R. G. AND A. A. EKDALE. 1984. Trace Fossil Preservation in Flint in the European Chalk. Journal of Paleontology, 58:298–311.

BROMLEY, R. G. AND A. A. EKDALE. 1986. Composite ichnofabrics and tiering of burrows. Geology Magazine, 123:59–65.

BROMLEY, R. G. 1996. Trace Fossils: Biology, Taphonomy and Applications (second edition). New York, Chapman and Hall, 361p.

CAÑAS, F. 1995. Early Ordovician carbonate platform facies of the Argentine Precordillera: restricted shelf to open platform evolution, p. 221–224. *In* J. D. Cooper, M. L. Droser, and S. C. Finney (eds.), Ordovician Odyssey: Short papers for the Seventh International Symposium on the Ordovician System. The Pacific Section Society for Sedimentary Geology Book 77.

CARMONA, N. B., L. A. BUATOIS AND M. G. MANGANO. 2004. The trace fossil record of burrowing decapod crustaceans: evaluating evolutionary radiations and behavioral convergence. Fossils and Strata, 51:141–153.

CLAYTON, C. J. 1986. The chemical environment of flint formation in Upper Cretaceous chalks, p. 43–54. *In* G. D.Sieveking, and M. B. Hart (eds.), The Scientific Study of Flint and Chert. Cambridge University Press, Cambridge.

CRIMES, T. P. 1974. Colonization of the early floor. Nature, 248:328–330.

DICKINSON, W. R. 1992. Cordilleran sedimentary assemblages, p.417–422. *In* B. C. Burchfiel, P. W. Lipman, and M. L. Zoback, (eds.), The Cordilleran Orogen: Conterminous U.S.: Boulder, Colorado, Geological Society of America. The Geology of North America, G-3.

DROSER, M. L. AND D. J. BOTTJER. 1986. A semiquantitative field classification of ichnofabric. Journal of Sedimentary Research, 56:558–559.

DROSER, M. L. AND D. J. BOTTJER. 1988. Trends in depth and extent of bioturbation in Cambrian carbonate marine environments, western United States. Geology, 16:233–236.

EKDALE, A. A. 1985. Paleoecology of the marine endobenthos. Palaeogeography, Palaeoclimatology, Palaeocology, 50:63-81.

EKDALE, A. A. 1988. Pitfalls of paleobathymetric interpretations based on trace fossil assemblages. Palaios, 3:464–472.

EKDALE, A. A. AND R. G. BROMLEY. 1984. Comparative Ichnology of Shelf-Sea and Deep-Sea Chalk. Journal of Paleontology, 58:322–332.

EKDALE, A. A., R. G. BROMLEY AND S. G. PEMBERTON. 1984. Ichnology: trace fossil in sedimentology and stratigraphy. Tulsa, Oklahoma, Society of Economic Paleontologists and Mineralogists: Short Course No. 15, 317 p.

EKDALE, A.A. AND R.G. BROMLEY 1991. Analysis of composite ichnofabrics: An example in Uppermost Cretaceous chalk of Denmark. Palaios, 6:232–249.

FELDMANN, R. M. AND C. E. SCHWEITZER. 2010. The oldest shrimp (Devonian: Famennian) and remarkable preservation of soft tissue. Journal of Crustacean Biology, 30:629–635.

HABERLAH, D., C. STRONG, D. PIRRIE, G. K. ROLLINSON, P.GOTTLIEB, W. S. K. BOTHA, AND A. R. BUTCHER. 2011. Automated petrography applications in Quaternary Science. Quaternary Australasia, 28:3–12.

HINTZE, L. F., G. C. WILLIS, D. Y. M. LAES, D. A. SPRINKEL, AND K. D. BROWN. 2000. Digital Geologic Map of Utah. Utah Geologic Survey.

LEHANE, J. R., AND A. A. EKDALE. 2013. Fractal analysis of graphoglyptid trace fossils. *Palaios*, 28:23-32.

LEHANE, J. R., AND A. A. EKDALE. 2014. Analytic tools for quantifying the morphology of invertebrate trace fossils. Journal of Paleontology, 88:747–759.

MAGWOOD, J. AND A. A. EKDALE. 1994. Computer aided analysis of complex ichnofabrics in deep-sea sediments. Palaios, 9(1):102–115.

MÁNGANO, M. G., AND L. A. BUATOIS. 2007. Trace fossils in evolutionary paleoecology, p. 391–409. *In* W. Miller III (ed.), Trace Fossils: Concepts, Problems, Prospects. Amsterdam, Elsevier.

MATTES, B. W. AND E. W. MOUNTJOY. 1980. Burial dolomitization of Upper Devonian Miette buildup, Jasper National Park, Alberta. Society of Economic

Paleontologists and Mineralogists Special Publication, 28:259–297.

MILLER, J. F., K. R. EVANS AND B. F. DATTILO. 2012. The great American carbonate bank in the miogeocline of western central Utah: Tectonic influences on sedimentation, p. 769−854. *In* J. R. Derby, R. D. Fritz, S. A. Longacre, W. A. Morgan, and C. A. Sternbach (eds.), The great American carbonate bank: The geology and economic resources of the Cambrian−Ordovician Sauk megasequence of Laurentia. American Association of Petroleum Geologists Memoir 98.

MILLER, M. F. AND S. E. SMAIL. 1997. The semiquantitative method for evaluating bioturbation on bedding planes. Palaios, 12:391–396.

MYROW, P. M. 1995. Thalassinoides and the enigma of Early Paleozoic open-framework burrow systems. Society for Sedimentary Geology, 10:58-74.

ORR, P.J. AND D. E. G. BRIGGS. 1999. Exceptionally preserved conchostracans and other crustaceans from the Upper Carboniferous of Ireland. *Special Papers in Palaeontology,* **62:68**.

PEMBERTON, S. G., AND FREY, R. W. 1985. The *Glossifungites* Ichnofacies: modern examples from the Georgia coast, USA, p. 237-259. *In* H. A. Curran (ed.), Biogenic structures: their use in interpreting depositional environments: Tulsa, Oklahoma, SEPM Special Publication 35.

PEMBERTON, S.G., J. A. MACEACHERN AND R. W. FREY. 1992. Trace fossil facies models: their environmental and allostratigraphic significance, p. 47–72. *In* Walker, R.G., and James, N.P. (eds.), Facies Models: A Response to Sea-Level Changes. Geological Association of Canada.

POOLE, F. G. 1992. Latest Precambrian to latest Devonian time; Development of a continental margin, p.46. *In* B. C. Burchfiel, P. W. Lipman, and M. L. Zoback, (eds.), The Cordilleran Orogen: Conterminous U.S.: Boulder, Colorado, Geological Society of America. The Geology of North America, G-3.

ROBINSON, R. A. 1960. Lower and Middle Cambrian stratigraphy of the eastern Great Basin, p. 43–52. *In* J. W. Boettcher and W. W. Sloan (eds.), Guidebook to the geology of east-central Nevada: Intermountain Association of Petroleum Geologists. Eleventh Annual Field Conference Guidebook.

SALTZMAN, M. R. 2005. Phosphorus, nitrogen, and the redox evolution of the Paleozoic oceans. Geology, 33:573–576.

SCHMIDT, F. 1986. Flint stratigraphy and its relationship to archeoecology, p. 1–5. *In* G. de C. Sieveking and M. B. Hart (eds.), The Scientific study of flint and chert. Cambridge University Press, Cambridge, England.

SEARS, J. W. AND R. A. PRICE. 2003. Tightening the Siberian connection to western Laurentia. Geological Society of America Bulletin, 115:943–953.

SEILACHER, A. 1953. Uber die Methoden der Palichnologie, 1, Studien von Palichnologie. Neues Jahrbuch für Geologie und Palaontologie, Abhandlungen, 96:421–52.

SEILACHER, A. 1967a. Bathymetry of Trace Fossils. Marine Geology, 5:413–428.

SEILACHER, A. 1967b. Fossil behavior. Scientific American, 217:72–80.

SEILACHER, A. 2007. Trace fossil analysis. Springer-Verlag, Berlin, Heidelberg, New York, 226p.

SHAPIRO, R. S. 2007. Stromatolites: A 3.5 Billion-Year Ichnologic Record, p. 382–390. *In* W. Miller III (ed.), Trace Fossils: Concepts, Problems, Prospects. Elsevier, Amsterdam.

SHEEHAN, P. M. AND D. R. J. SCHIEFELBEIN. 1984. The Fossil Record Thalassinoides, from the Upper Ordovician of the Great Basin: deep burrowing in the Early Paleozoic. Journal of Paleontology, 58:(2) 440–447.

STOKES, W. L. 1986. Geology of Utah. Salt Lake City, Utah, Utah Museum of Natural History, University of Utah and Utah Geologic and Mineral Survey, Department of Natural Resources, 47 p.

TAYLOR, M. E. AND H. E. COOK. 1976. Continental shelf and slope facies in the Upper Cambrian and Lower Ordovician of Nevada. Brigham Young University Geology Studies, 23(2):181–214.

THULBORN, R. A. 1990. Dinosaur traces. London, Chapman and Hall.

UCHMAN, A. 1995. Taxonomy and paleoecology of flysh trace fossil: the Marnoso-arenacea Formation and associated facies (Miocene, Northern Apennines, Italy). Beringeria, 15:1–115.

VASCONCELOS C., J. A. MCKENZIE, S. BERNASCONI, D. GRUJIC, AND A. J. TIEN. 1995. *Microbial mediation as a possible mechanism for natural dolomite formation at low temperatures. Nature, 337:220–222.*

VENDRASCO, M. J., AND B. RUNNEGAR. 2004. Late Cambrian and Early Ordovician stem group chitons (Mollusca: Polyplacophora) from Utah and Missouri. Journal of Paleontology, 78(4):675–689.

WHITAKER, F. F AND Y XIAO. 2010. Reactive transport modeling of early burial dolomitization of carbonate platforms by geothermal convection. American Association of Petroleum Geologists Bulletin, 94(6): 889–917.

CHAPTER 3

PALEOECOLOGY OF UPPER CAMBRIAN CARBONATE UNITS

IN THE HOUSE RANGE AND SOUTH LAKESIDE

MOUNTAINS, UTAH

3.1 Abstract

Upper Cambrian sedimentary sequences in the House Range and South Lakeside

Mountains in the Great Basin of western Utah are dominated by marine carbonate facies.

Burrowed, stromatolitic, dolomitized limestones are overlain by parallel-bedded

carbonate strata with no burrows or stromatolites in both field areas. A regressive

sequence was responsible for the sedimentological characteristics of the overlying strata,

which included the Red Tops Member of the Notch Peak Formation in the southern

House Range and the Garden City Formation in the South Lakeside Mountains. A global

negative carbon excursion event (HERB) is recorded in the both the Red Tops Member

and the Garden City Formation.

Abundant stromatolites were important for shelter and food for the limited variety

of benthic organisms in the Late Cambrian. The Hellnmaria and Upper Carbonate

Members contain cyanobacterial stromatolite facies containing *Thalassinoides*

paradoxicus burrows inside and surrounding the stromatolites. Primary producers in the

ecosystem were photosynthetic cyanobacteria, and since there were no multicellular

plants of any kind, these small prokaryotes formed the base of the trophic structure of the

benthic community. The primary consumers in both field areas were herbivorous

mollusks, *Matherella walcotti* (gastropod) and *Matherella wahwahensis*

(polyplacophoran), as well as the unknown organism that created the *Thalassinoides* in

the stromatolites. *Thalassinoides* paleoecology is key to understanding the

paleoenvironment during the Late Cambrian and the stromatolite facies that they lived in

is a crucial part of the paleoecology and paleoclimate at this time.

3.2 Introduction

   The field areas of this study are located in the Great Basin of western Utah, and

the rock units include both the Hellnmaria Member of the Notch Peak Formation in the

southern House Range and the Upper Carbonate Member of the St. Charles Formation in

the South Lakeside Mountains. These two locations have correlative Upper Cambrian

strata representing a shallow, tropical, quiet water paleoenvironment (Figure 3.1).

   Global sea level was high during the Late Cambrian, and there were no polar

icecaps. The Cambrian was a time of higher atmospheric temperatures, which would have

facilitated limestone deposition (Viezer et al., 1999). There is evidence of greenhouse

conditions during this time, and because of the high sea level during the Late Cambrian,

there were extensive carbonate platforms extending hundreds of miles (Sheehan and

Schiefelbein, 1984). These platform edges kept the wave energy low enough so that

stromatolites and ooids were not ripped apart within the platform interior.

   Abundant stromatolites in both units include laterally linked hemispheroids,

stacked hemispheroids, and columnar structures. *Thalassinoides* burrows commonly are

found in quiet water carbonates deposited near or below wave base. *Thalassinoides paradoxicus* occurs profusely in the Hellnmaria and Upper Carbonate Members and exhibits a close association with stromatolites, and the same trace fossils do not occur in non-stromatolitic facies of the same age elsewhere.

This study demonstrates the importance of trace fossils in the interpretation of paleoecology, paleoclimate, and paleoenvironment of the ancient carbonate strata during the Late Cambrian in the Great Basin. Trace fossils are particularly useful, because much of the strata of this age and region lack a substantial body fossil record.

3.3 Cambrian Setting

During the Cambrian the continents of Siberia and Laurentia had split along a rifting margin. This was a time of higher sea level, which fell shortly thereafter. The Red Tops Member, which overlays the Hellnmaria Member in the House Range, was a lowstand interval when the sea level fell. This dramatic change in sea level greatly influenced marine life at this time and at the location where these strata were deposited.

The Early and Middle Cambrian were intervals of increasing faunal diversity, with cyanobacteria, trilobites, echinoderms, mollusks, and brachiopods as dominant marine organisms. Typical reef builders such as corals, stromatoporids, and bryozoans were absent from the Cambrian seas. Stage 10 of the Cambrian is a very significant time unit in evolutionary history, because it was a time when family and genus level diversity of organisms in the marine realm decreased markedly (Sepkoski, 1981, 1997).

During the Late Cambrian, the most important primary producers were photosynthetic algae, mostly cyanobacteria. Since there were no multicellular plants,

such as trees or grasses of any kind, these small prokaryotes formed the base of the trophic structure. The primary consumers in both field areas were *Matherella walcotti*, *Matherella wahwahensis*, and the organisms that created the *Thalassinoides* in the stromatolites. Usually in Late Cambrian strata there is evidence of some secondary consumers (i.e., predators) such as anomalicarids (Vendrasco and Runnegar, 2004), but the strata in this field area preserve no such predators (Figure 3.2).

The Upper Cambrian rocks in this field study are mostly dolomitized, making it difficult to determine all the possible organisms. The extensive recrystallization of the limestone to dolomite has destroyed some of the internal structure and fossils. This is one reason why the *Thalassinoides* in the stromatolites are significant. They are pervasive and are visible in the Upper Cambrian stromatolite facies in both the northern and southern Upper Cambrian exposures at both study areas of the Great Basin.

3.4 Stromatolite Paleoecology

Stromatolites are biogenic structures that record the interaction between microbial communities and sediments (Shapiro, 2007). The Cambrian stromatolites are sediment-generated structures made from the accumulation of successive mats of cyanobacteria filaments. Stromatolites are considered to be laminated microbial structures, whereas thrombolites are unlaminated microbial structures (Shapiro, 2007), both of which are exposed in the rocks in this study. Layers are produced when calcium carbonate is precipitated onto the algal mat along with other sediments. Precipitation is initiated when cyanobacteria depletes carbon dioxide in water, making it less acidic. Stromatolites yield information about behavior and environmental constraints of microbial ecosystems.

Stromatolites from the Late Cambrian and Early Ordovician found in the House Range,
Wah Wah Range, and South Lakeside Mountains have *Thalassinoides paradoxicus*
burrows inside many of them.

3.5 *Thalassinoides* Feeding Patterns

Thalassinoides is a highly branched burrow system. The burrow tunnels branch in
a "Y" or "T" pattern generally. The burrows observed in this study show similar angles,
which are close in degrees to one and other, indicating travel in all directions. The
burrowing organisms would have been able to move in multiple directions throughout the
entire burrow system using a range of differing pathways. If the angles between burrows
create burrow tunnels that are more parallel to each other, that would mean the organism
was using different burrow tunnels to travel in almost the same direction, and this would
be inefficient, so this is not a pattern usually observed in nature (Lehane and Ekdale,
2014).

3.6 Possible *Thalassinoides* Architects in the Late Cambrian

Thalassinoides burrows are mostly thought to be made by decapod crustaceans
(Ekdale 1984), but there is no evidence of such organisms living in the Late Cambrian.
Decapod crustaceans have flexible, highly mobile bodies with lots of specialized
appendages, commonly including robust claws used in digging. In modern environments,
various decapod taxa create a large diversity of branching and anastomosing burrows
systems, providing direct analogues for some types of ancient *Thalassinoides* (Bromley,
1996). Decapods today have a wide variety of feeding habits, including deposit-feeding,

suspension-feeding, and predation and scavenging. Because the geologic record of decapod body fossils begins in the Middle Paleozoic, decapod crustaceans could not have been the trace makers of *Thalassinoides* in this study.

The entire body fossil record of decapod crustaceans is extremely sparse before the Triassic (Feldmann and Schweitzer, 2010). The earliest body fossils of decapods are about 120 million years younger than the Hellnmaria Member, where the *Thalassinoides* were observed in this study. Some Cambrian crustaceans, such as phyllocarids and anomalicarids, had flexible, highly mobile bodies with lots of appendages. They may have included deposit-feeders, suspension-feeders, and/or predators. But *Thalassinoides* burrows have not been attributed to either of these organisms. They were present during the Late Cambrian, but they typically are not well preserved, and their body fossils have not been found in association with Cambrian *Thalassinoides* occurrences anywhere, including in the strata studied for this thesis.

Trilobites have a flattened body with multiple pairs of chitinous, biramous appendages located beneath a calcite, segmented carapace. They have been known to produce a large diversity of trace fossils, including shallow pits and gently sinuous and broadly looping trails (Seilacher, 1970). Most trilobites probably were deposit-feeders, and they were quite abundant in Cambrian rocks, but they are rarely found in association with Lower Paleozoic *Thalassinoides* in Utah (Pickerill, 1995).

Annelids have elongate, flexible bodies, which provide maximum agility, and they can produce a large diversity of burrow geometries such as *Skolithos*, *Planolites*, *Arenicolites*, *Diplocraterion*, and possibly *Zoophycos*. They also have a wide variety of feeding habits, including deposit feeding. Annelid trace fossils are abundant in the Upper

Cambrian rocks, but they have never been known to make anastomosing burrow systems (Mángano and Buatois, 2007).

The most common macrofossils in the stromatolitic facies in the two study areas are mollusks. Archeogastropods (*Matherella wahwahensis*) occur in close association with *Thalassinoides* in Upper Cambrian stromatolitic facies. They are found on top of the stromatolites only, appearing to have been epifaunal grazers on the stromatolites. Polyplacophores (*Matthevia walcotti*) also occur in the *Thalassinoides*-rich stromatolitic facies in this study. They also appear to have been epifaunal grazers on the stromatolites. This may be supporting evidence that *Matherella* and *Matthevia* were not deep burrowers and therefore were not capable of creating a network of anastomosing burrows. *Matthevia* were polyplacophorans (chitons) with eight rigid, conical plates running along the top of their body, which allowed flexing of the body so that the animal could curl up into a ball. They could move their bodies up and down, but *Matthevia* could not easily move side to side in the type of direction required to create *Thalassinoides*.

Gastropods and polyplacophorans both have shells of aragonite. This mineral is slightly more soluble than calcite, supporting the idea that the dissolution of the original shell material probably occurred soon after burial (Vendrasco and Runnegar, 2004). If the *Matherella* and *Matthevia* shells dissolved soon after burial, and the *Thalassinoides* burrows were heavily bioturbated sediment, then there could be a connection between available void spaces and the propensity for both the burrows and the *Matterella* and *Matthevia* shells to become silicified.

Some of the earliest pelecypods can be found in the Upper Cambrian, but they were nonsiphonate and therefore could not have been deep infaunal burrowers. Lingulid

(inarticulate) brachiopods have a bivalved shell with a long pedicle, which they used for burrowing, but they are exclusively suspension-feeders that make simple vertical shafts. In addition, lingulids have never been found in association with Cambro-Ordovician *Thalassinoides*. Primitive echinoderms, such as eocrinoids and helicoplacoids, are known from the Cambrian of western North America, but they are known to be epifaunal.

Conodonts also have been discovered in these sediments (Miller et al., 2006; Miller et al., 2011; Miller et al., 2012). The conodont organisms were mostly soft-bodied, motile, eel-like animals, but it is unknown if the conodont animals were grazers, burrowers, or predators.

3.7 Shallow Marine Environment

The study sites in the House Range and South Lakeside Mountains both contain burrowed stromatolites. Burrowing organisms thrive in open marine environments, and stromatolites are generally found in restricted marine environments within the photic zone, because they need photosynthesis to grow. The carbonate units observed in both field areas show an ichnofabric index of *ii*3 to *ii*5 (moderate to thorough bioturbation). *Thalassinoides* in the modern oceans, as well as in Upper Cambrian rock units of the Great Basin, commonly are associated with a firmground substrate in lime mud, which is prevalent in shallow marine environments. At the Upper Cambrian study sites, wave energy probably was low to moderate with shallow, clear water, which is required for photosynthesis by the cyanobacteria.

3.8 Paleoclimate

Evidence of the paleoclimate during the Upper Cambrian in the Great Basin of Utah is determined in part by the paleogeographic location of the field sites during this time, as well as by carbon isotope excursion events, depositional environment, and trace fossil evidence. Paleolatitude of the field area in this study was approximately 5°S, so the setting clearly was tropical.

The study area during the Late Cambrian was situated on a passive marine margin, which was a shallow-marine platform characterized by widespread deposition of limestones. Carbonate precipitation and photosynthesis by benthic organisms removed a significant amount of the carbon dioxide from the atmosphere, fixing it as carbonate in limestone, which was being deposited in the study area during the Late Cambrian (Saltzman et al., 2005).

3.9 Carbon Excursions in Upper Cambrian Units in Western Central Utah

The low point of the Sauk sea level regression occurred approximately 499 Ma years ago, and it coincided with the peak of the Steptoean Positive Carbon Isotope Excursion (SPICE). Slightly below this level in both study areas is a bed with paleokarst features representing this regression lowstand (Miller, 2011). The karsted section lies stratigraphically below the stromatolitic facies of this study, and it shows how the rock record reflects the SPICE event in the field areas.

The name "HERB" refers to a negative carbon excursion that preceded the SPICE event (Figure 3.3), but the name HERB is not an acronym (Ripperdan, 2002). The HERB isotopic event is marked by a negative $\delta^{13}C$ anomaly, which has been identified in

western central Utah, Australia, Newfoundland, China, and western North America (Figure 3.4) (Miller, 2011; Ripperdan et al., 1992; Ripperdan, 2002;). This pronounced negative $\delta^{13}C$ anomaly is also evident in Stage 10 Upper Cambrian carbonates in Argentina in the La Flecha (Sial et al., 2008). The HERB occurred in Stage 10 of the Upper Cambrian of the Great Basin, which is located stratigraphically at the Hellnmaria-Red Tops boundary (Figure 3.5), when the carbon isotope values begin to shift toward a large amplitude negative peak occurring in the Red Tops Member (Ripperdan et al., 1992).

The HERB event occurred very shortly after the deposition of the stromatolite facies, which is the focus of this study. Parallel strata truncated the uppermost section of *Thalassinoides*-burrowed stromatolites with no burrows at approximately the same time as the suggested time of the HERB event. This event also correlates with a significant drop in sea level that is recorded in the Red Tops Member, which overlies the rocks in this study. The boundary between the Hellnmaria and Red Tops Member is located 3 m above the top of the Hellnmaria Member exposure at the field area at Steamboat Pass in the southern House Range, where samples were collected for this study. The stromatolites and the burrows do not appear again in either field area until right after this sea level regressive event. Therefore, the study area appears to have been a restricted marine platform with high sea level just prior to the HERB. A regressive sequence was responsible for the characteristics of the Red Tops Member, which contains no burrows or stromatolites in the field area at the boundary between the Hellnmaria and Red Tops members.

3.10 Sea Level Changes

During the HERB event, there was sea level regression, and the stromatolitic

facies is not present in the overlying strata of the Red Tops Member at the contact or the

correlative strata overlaying the Upper Carbonate Member in the South Lakeside

Mountain study site. Stromatolites do appear higher in section in the Red Tops Member,

however. This could be possible evidence of a decrease of primary productivity as a

possible cause or contributor to the negative $\delta^{13}C$ anomaly if this occurred on a global

scale.

In Argentina, there is also a negative carbon isotope excursion recorded in the La

Flecha Formation carbonates in Stage 10 of the Upper Cambrian, and this event is likely

equivalent to that registered in Stage 10 carbonates of North America and Australia, thus

suggesting the possibility of a global event (Sial et al., 2008).

3.11 Late Cambrian Extinction Event

It is possible that the HERB event caused a loss of primary productivity leading to

a possible extinction event at the end of the Cambrian (Riding, 2006; Ripperdan, 2002).

During the Cambrian, primary productivity is attributed mostly (if not entirely) to algae

and cyanobacteria, and the loss of primary productivity may have facilitated a more

negative carbon shift (Riding, 2006). There was a decrease in diversity at the end of the

Cambrian (Alroy et al., 2008; Sepkoski, 1997). There was very little variety of

photosynthetic primary producers during this time, since there was little diversity of

marine algae and no plant life on land.

Cyanobacteria were principally responsible for the oxygen levels in Earth's atmosphere and oceans during this time, and there were extremely high levels of carbon in the atmosphere relative to modern levels (Veizer et al., 1999). The Late Cambrian also was a time of elevated sea level, approximately as high as it was during the Cretaceous. Because the situation was quite different from sea level today, it is difficult to use modern analogues when investigating carbon isotope excursions during the Cambrian, as the overall environment was vastly different.

3.12 Shoreline Regression

Shoreline regression, microbial activity, and carbonate platform exposure are some other possible causes of negative carbon excursions in the geologic record (Ripperdan, 2002). Weathering and erosion result from carbonate platform exposure, and they may have provided a source of organic carbon facilitating the negative carbon isotopic excursion in the Great Basin.

The Red Tops Member was deposited during a fall in sea level, while prior to this time, the Hellnmaria Member and the Upper St. Charles Formation were deposited during a transgressive period with a rise in sea level. A shoreline regression then may have been the result of climatic cooling and glaciation. A cooler climate would preclude organisms that are cold intolerant. Lower oxygen levels, which were a result of reduced primary productivity, could have led to extinction of marine taxa. These events occurring together could have created marine $\delta^{13}C$ negative values. The Late Cambrian climate in this region was therefore a warm tropical paleoclimate, based on (a) low paleolatitude, (b) widespread carbonate precipitation, and (c) a high $\delta^{13}C$ signature.

3.13 Conclusion

The Hellnmaria Member and the Upper Carbonate Members are correlative Upper Cambrian strata in the southern House Range and South Lakeside Mountains of western Utah, respectively. Both members represent a burrowed stromatolite facies where the probable burrowers most likely were mining shallow marine silts or muds for food as well as seeking protection from predators in the complex web of burrows. The burrows (*Thalassinoides paradoxicus*) were excavated originally as open burrows that subsequently were filled with sediment of a different color from that of the surrounding sediment. The substrate had an effect on how these trace fossils were created and preserved, and the burrow architect must have lived in relatively firm sediment, since the burrow wall shows no signs of reinforcement. As a result of dolomitization and deformation due to age and burial of the strata, many diagnostic features, such as uniform tunnel diameters, regular branching angles, and bioglyphs, are absent from these Upper Cambrian *Thalassinoides*.

Abundant stromatolites were important for shelter and food for a variety of organisms in the Late Cambrian strata that are the focus of this study. The most important primary producer was photosynthetic algae, mostly cyanobacteria. Because there were no multicellular plants, these small prokaryotes formed the base of the trophic structure. The primary consumers in both field areas were *Matthevia walcotti*, *Matherella wahwahensis*, and the unknown organism that created the *Thalassinoides* in the stromatolite facies of the Hellnmaria and Upper Carbonate Members.

The Hellnmaria and Upper Carbonate Members represent a paleoenvironment in

shallow, tropical, quiet water, perhaps in a highly alkaline setting. The *Thalassinoides* in this study are unlined and not reinforced in any way, so they would have required a calm, shallow, firmground environment, or else the burrows would have collapsed immediately due to shifting or unpackable sediments.

Interbedded lime mudstone and grainstone is common in the Upper Carbonate Member as well as in the Hellnmaria Member. This setting represents an open marine platform with a platform edge lying to the west. The platform edge may have kept the energy low enough for stromatolites and ooids to form in the interior of the platform and not be ripped apart by larger waves or storm waves. The *Thalassinoides* exhibit a close association with stromatolites and are not found in non-stromatolitic facies elsewhere.

High sea levels were present when the *Thalassinoides* producers burrowed the Late Cambrian stromatolites. Sea level correlates with possible rapid sea floor spreading, which correlates also with higher carbon dioxide levels in the atmosphere, possibly due to increased rates of carbon dioxide outgassing. All of these scenarios would lead to higher atmospheric temperatures, facilitating limestone deposition. This is evidence of greenhouse conditions that existed during this time.

A global negative carbon excursion event (HERB) is recorded in the strata just above the Hellnmaria and Upper Carbonate members in both field areas. Parallel strata with no burrows or stromatolites truncate the uppermost part of the section with *Thalassinoides*-burrowed stromatolites at approximately the same time as the suggested time of the HERB event. Both members are overlain by sediments with no stromatolites or trace fossils at the rock unit boundary, which reflects the transition from highstand to lowstand conditions. A regressive sequence influenced the sedimentologic characteristics

of the overlying strata, which included the Red Tops Member in the southern House Range and the Garden City Formation in the South Lakeside Mountains.

This study demonstrates the importance of trace fossils in the interpretation of paleoecology, paleoclimate, and depositional environments of the ancient carbonate strata during the Late Cambrian in the Great Basin. This is particularly useful, because most strata of this age in this region lack a substantial body fossil record, which otherwise would aid in the paleoecologic interpretation of these interesting rocks.
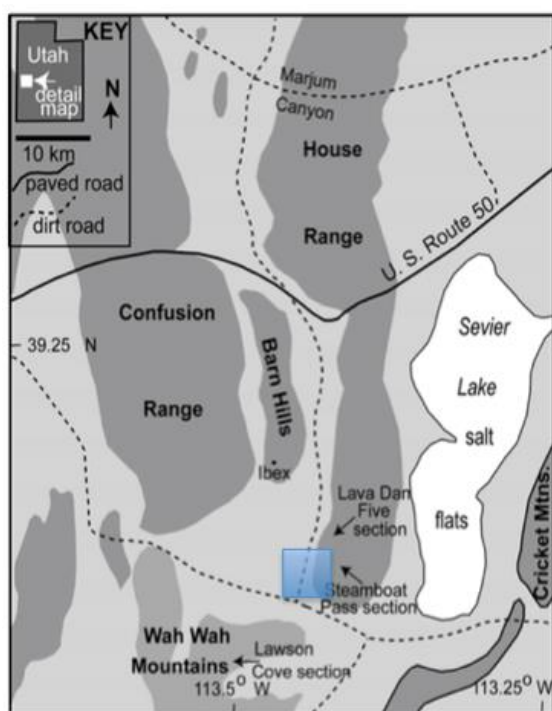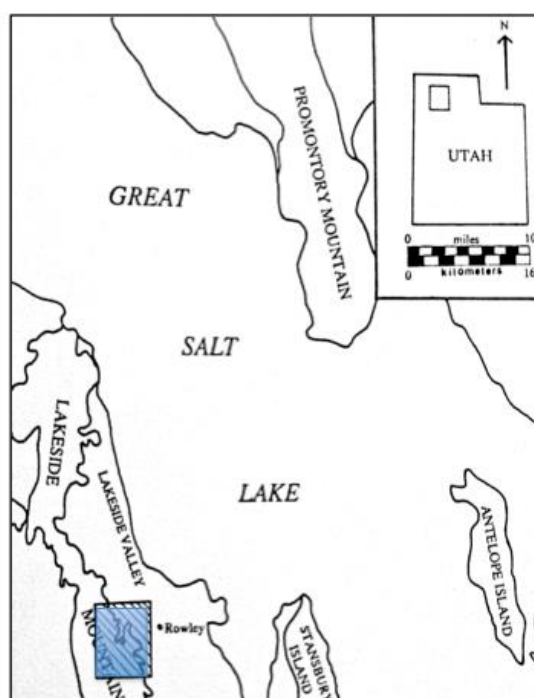
Figure 3.1. Maps of both research areas in the southern House Range and South Lakeside Mountains in the Great Basin of western Utah; the blue shaded areas are the study sites (modified after Landing et al., 2011).

Late Cambrian cyanobacteria stromatolites

Mollusk grazers
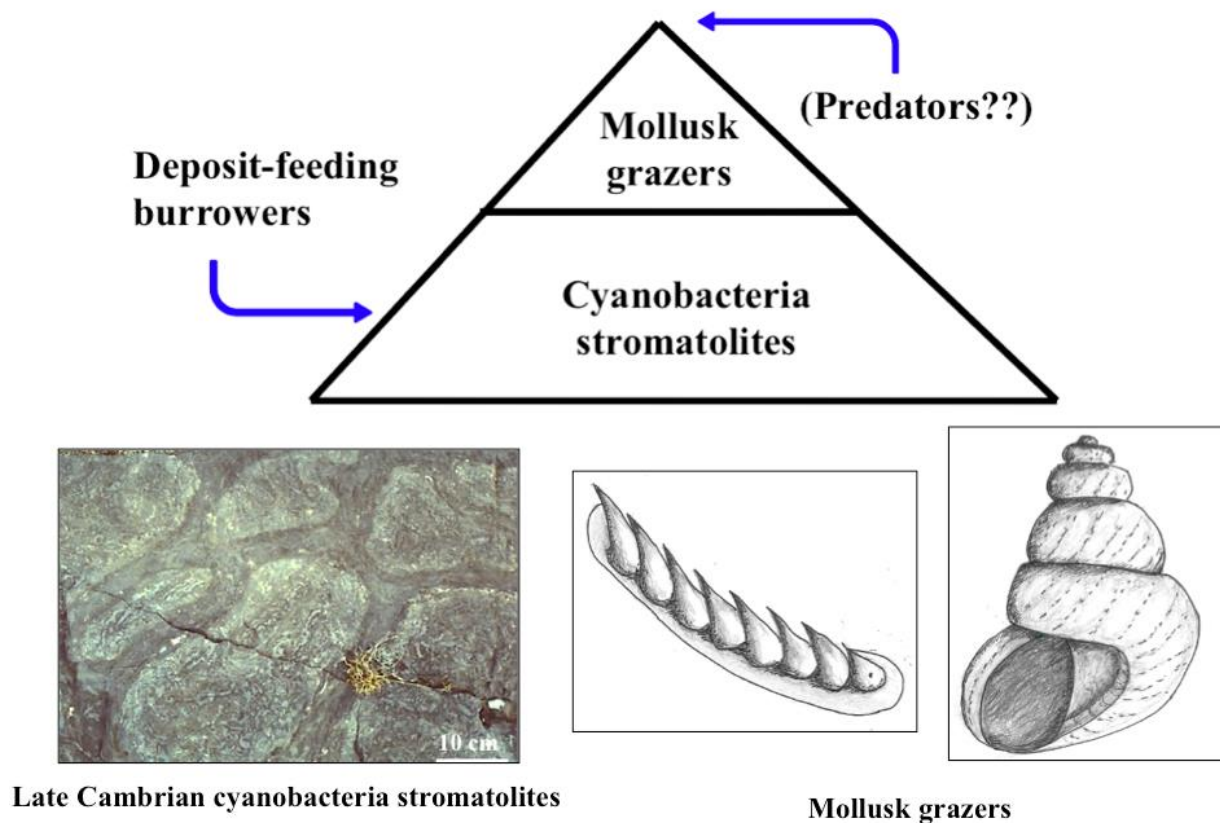
Figure 3.2. Diagram showing the generalized trophic structure of the benthic communities along the carbonate belt where the stromatolitic facies formed in the Late Cambrian. Primary producers (cyanobacteria) and primary consumers (molluscan grazers and unidentified infaunal deposit feeders) are represented in the trophic pyramid, but secondary consumers (predators) apparently are unrepresented in the fossil association, because no fossils of obvious predators have been found in the study sites.
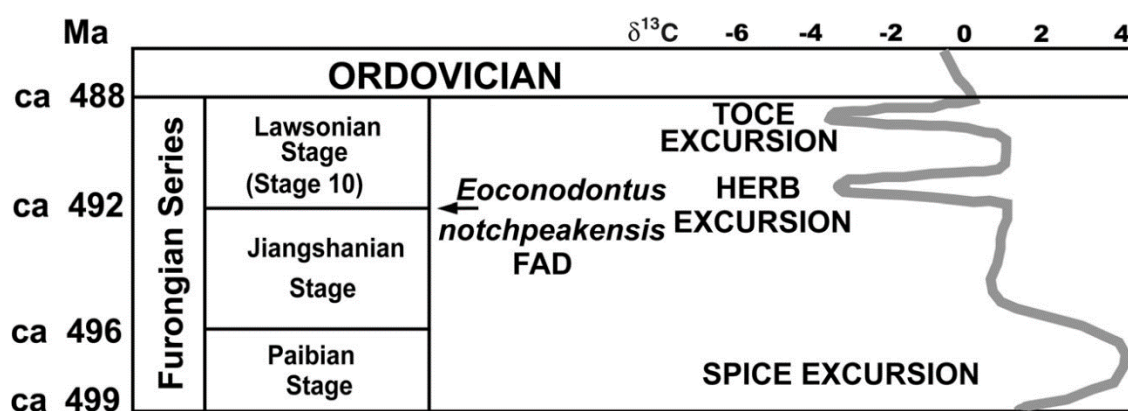
Figure 3.3. Pronounced negative ^{13}C isotope excursion (−4‰) recorded in the Upper Cambrian strata of the Great Basin (modified from; Miller, 1995; Ripperdan 1992). *Eoconodontus notchpeakensis* (a conodont) is the key index fossil in this sequence.
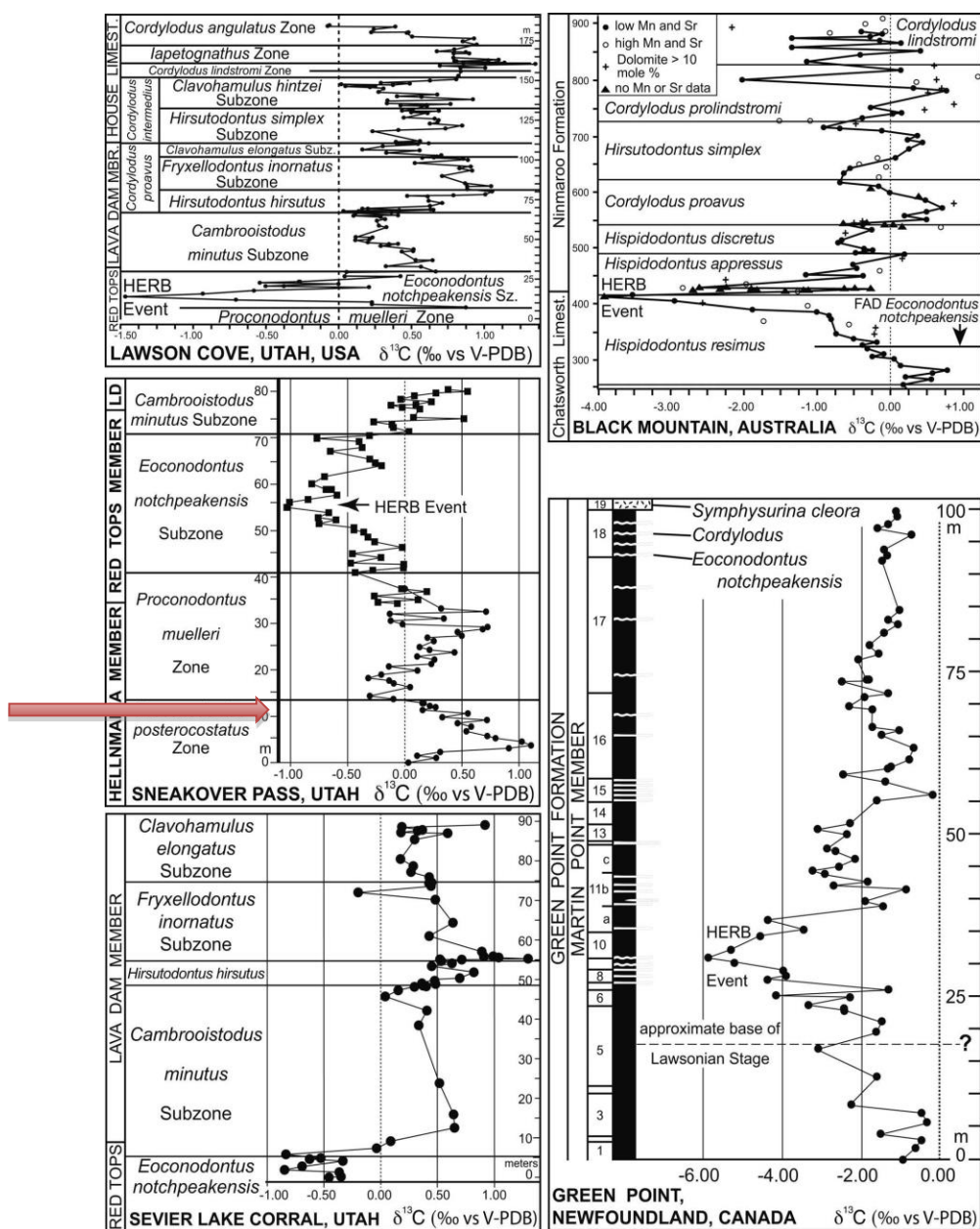
Figure 3.4 Diagram of the HERB ^{13}C isotopic excursion that has been recognized in Utah, Australia, Newfoundland, and China; the red arrow marks the location of this study in Utah (modified after Miller et al., 2011).
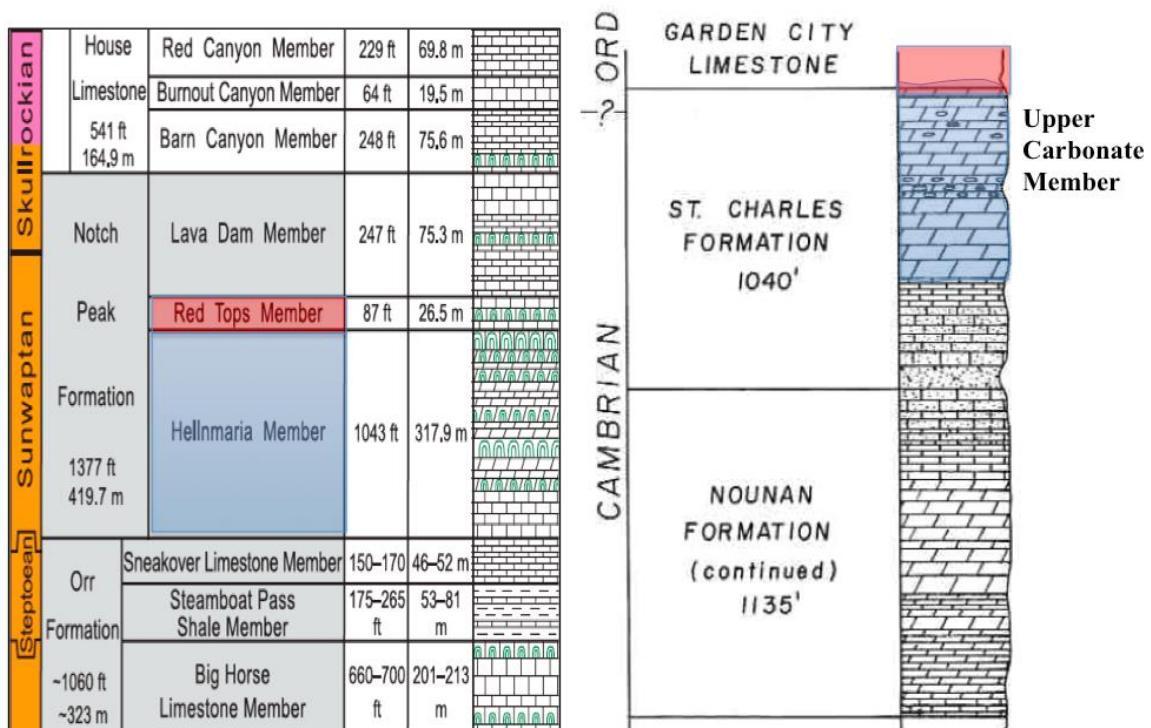
Figure 3.5. Stratigraphic columns of the southern House Range (left) and the South Lakeside Mountain (right). The blue shaded sections indicate the members that are the focus of this study, and the red shaded areas are the approximate positions of sediments deposited during the HERB negative carbon excursion (modified after Miller et al., 2011; Miller, 2012; Stokes 1986).

3.14 References

ALROY, J., D. J. BOTTJER, M. FOOTE, F. T. FURSICH, P. J. HARRIES, A. J. W. HENDY, S. M. HOLLAND, L. C. IVANY, W. KIESSLING, M. A. KOSNIK, C. R. MARSHAL, A. J. MCGOWAN, A. I. MILLER, T. D. OLSZEWSKI, M. E. PATZKOWSKY, S. E. PETERS, L. VILLIER, P. J. WAGNER, N. BONUSO, P. S. BORKOW, B. BRENNEIS, M. E. CLAPHAM, L. M. FALL, C. A. FERFUSON, V. L. HANSON, A. Z. KRUG, K. M. LAYOU, E. H. LECKEY, S. NURNBERG, C. M. POWERS, J. A. SESSA, C. SIMPSON, A. TOMAOVCH AND C. C. VISAGGI. 2008. Phanerozoic Trends in the Global Diversity of Marine Invertebrates. Science, 321:97–100.

BROMLEY, R. G. 1996. Trace Fossils: Biology, Taphonomy and Applications (second edition). Chapman and Hall, New York, 361 p.

EKDALE, A. A. 1984. Comparative ichnology of shelf-sea and deep-sea chalk. Journal of Paleontology, 58:322–332.

FELDMANN, R. M. AND C. E. SCHWEITZER. 2010. The oldest shrimp (Devonian: Famennian) and remarkable preservation of soft tissue. Journal of Crustacean Biology, 30:629–635.

HINTZE, L. F., G. C. WILLIS, D. Y. M. LAES, D. A. SPRINKEL AND K. D. BROWN. 2000. Digital Geologic Map of Utah. Utah Geologic Survey.

HINTZE, L. F. AND B. J. KOWALIS. 2009. Geologic History of Utah. Brigham Young University Geologic Studies, Special Publication 9:1–225.

LANDING, E., S. R. WESTROP AND J. M. ADRAIN. 2011. The Lawsonian Stage – the *Eoconodontus notchpeakensis* FAD and HERB carbon isotope excursion define a globally correlatable terminal Cambrian stage. Bulletin of Geosciences, 86(3):621–640.

LEHANE, J. R. AND A. A. EKDALE. 2014. Analytic tools for quantifying the morphology of invertebrate trace fossils. Journal of Paleontology, 88(4):747–759.

MÁNGANO, M. G., AND L. A. BUATOIS. 2007. Trace fossils in evolutionary paleoecology, p. 391–409. *In* W. Miller III (ed.), Trace Fossils: Concepts, Problems, Prospects. Elsevier Publishing Company, Amsterdam, The Netherlands Amsterdam.

MILLER, J. F., R. L. ETHINGTON, R. E. EVANS, L. E. HOLMER, J. D. LOCH, L. E. POPOV, J. E. REPETSKI, R. L. RIPPERDAN AND J. F. TAYLOR. 2006. Proposed stratotype for the base of the highest Cambrian stage at the first appearance datum of *Cordylodus andresi*, Lawson Cove section, Utah, USA. Palaeoworld, 15:384–405.

MILLER, J.F., K. R. EVANS, R. L. FREEMAN, R. L. RIPPERDAN AND J. F. TAYLOR. 2011.

Proposed stratotype for the base of the Lawsonian Stage (Cambrian Stage 10) at the First Appearance Datum of *Eoconodontus notchpeakensis* (Miller) in the House Range, Utah, USA. Bulletin of Geosciences, 86(3):595–620.

MILLER, J. F., K. R. EVANS AND B. F. DATTILO. 2012. The great American carbonate bank in the miogeocline of western central Utah: Tectonic influences on sedimentation, p. 769–854. *In* J. R. Derby, R. D. Fritz, S. A. Longacre, W. A. Morgan, and C. A. Sternbach, (eds.), The great American carbonate bank:The geology and economic resources of the Cambrian – Ordovician Sauk megasequence of Laurentia. American Association of Petroleum Geologist Memoir 98.

PICKERILL, R. K. 1995. Deep-water marine *Rusophycus* and *Cruziana* from the Ordovician Lotbiniere Formation of Quebec. Atlantic Journal, 31:103–108.

RIDING, R. 2006. Microbial carbonate abundance compared with fluctuations in metazoan diversity over geological time. Sedimentary Geology, 185:229–238.

RIPPERDAN, R. L., M. MARGARITZ, R. S. NICOLL AND J. S. SHERGOLD. 1992. Simultaneous changes in carbon isotopes, sea level, and conodont biozones within Cambrian–Ordovician boundary interval at Black Mountain, Australia. Geology, 20:1039–1042.

RIPPERDAN, R. L. AND J. F. MILLER. 1995. Carbon isotope ratios from the Cambrian Ordovician boundary section at Lawson Cove, Wah Wah Mountains, Utah, p.129–132. *In* J. D. Cooper (ed.), Ordovician odyssey: short papers for the Seventh International Symposium on the Ordovician System. SEPM Society for Sedimentary Geology Pacific Section 77.

RIPPERDAN, R. L. 2002. The HERB event: end of Cambrian carbon cycle paradigm? Geological Society of America Abstracts with Programs, 34(6):413.

ROBINSON, R. A. 1960. Lower and Middle Cambrian stratigraphy of the eastern Great Basin, p. 43–52. *In* J. W. Boettcher and W. W. Sloan (eds.), Guidebook to the geology of east-central Nevada: Intermountain Association of Petroleum Geologists, Eleventh Annual Field Conference Guidebook.

RUNNEGAR B., J. POJETA, M. E. TAYLOR AND D. COLLINS. 1979. New species of the Cambrian and Ordovician chitons *Matthevia* and Chelodes from Wisconsin and Queensland: evidence for the early history of polyplacophoran mollusks. Journal of Paleontology, 53(6):1374–1394.

SALTZMAN, M. R. 2005. Phosphorus, nitrogen, and the redox evolution of the Paleozoic oceans. Geology, 33(7):573–576.

SEILACHER, A. 1967a. Bathymetry of Trace Fossils: Marine Geology, 5:413–428.

SEILACHER, A. 1967b. Fossil behavior. Scientific American, 217:72–80.

SEILACHER, A. 1970. *Cruziana* stratigraphy of "non-fossiliferous" Paleozoic sandstones, p. 447–476. *In* T. P. Crimes and J. C. Harper (eds.), Trace Fossils. Geological Journal Special Issue 3.

SEPKOSKI, J. J. 1981. A factor analytic description of the marine fossil record. Paleobiology, 7:36–53.

SEPKOSKI, J. J. 1997. Biodiversity: past, present, and future. Journal of Paleontology, 71:533–539.

SHAPIRO, R. S. 2007. Stromatolites: A 3.5 Billion-Year Ichnologic Record, p. 382–390. *In* W. Miller III (ed.), Trace Fossils: Concepts, Problems, Prospects. Amsterdam, Elsevier.

SHEEHAN, P. M. AND D. R. J. SCHIEFELBEIN. 1984. The Fossil Record Thalassinoides, from the Upper Ordovician of the Great Basin: deep burrowing in the Early Paleozoic. Journal of Paleontology, 58:(2) 440–447.

SIAL, A.N., S. PERALTA, V. P. FERREIRA, A. J. TOSELLI, F. G. ACEÑOLAZA, M. A. PARADA, C. GAUCHER, R. N. ALONSO AND M. M. PIMENTEL. 2008. Upper Cambrian carbonate sequences of the Argentine Precordillera and the Steptoean C-Isotope positive excursion (SPICE). Gondwana Research, 13:437–452.

VEIZER, J., D. ALA, K. AZMY, P. BRUCKSCHEN, D. BUHL, F. BRUHN, G. A. F. CARDEN, A. DIENER, S. EBNETH, Y. GODDERIS, T. JASPER, C. KORTE, F. PAWELLEK, O. PODLAHA, AND H. STRAUSS. 1999. 87Sr/86Sr, d13C and d18O evolution of Phanerozoic seawater. Chemical Geology, 161:59–88.

VENDRASCO, M. J., AND B. RUNNEGAR. 2004. Late Cambrian and Early Ordovician stem group chitons (Mollusca: *Polyplacophora*) from Utah and Missouri. Journal of Paleontology, 78(4):675–689.

APPENDIX

SUPPLEMENTARY DATA FOR THIN

SECTION AND QEMSCAN

ANALYSIS
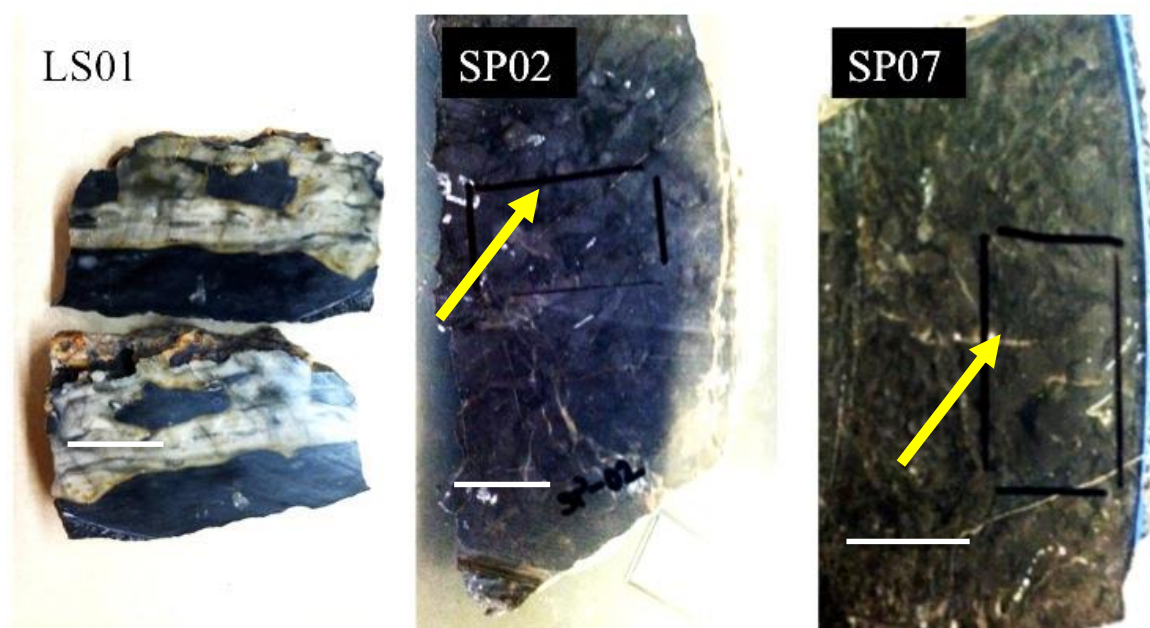
Figure A.1. Rock slabs used in Chapter 2 in Figure 2.11. LS01 is a polished serial cut section of dolomite containing silicified burrows. A thin section was made from this piece of rock. SP02 is a polished serial cut piece of dolomite that contains dolomite burrows, shown with the yellow arrow. A thin section was made of the section outlined on the rock with black marker. SP07 is a polished serial cut piece of a dolomitized stromatolite containing dolomite burrows. The black square marked with pen is where the thin section was made on this rock. Scale bars are 2cm.
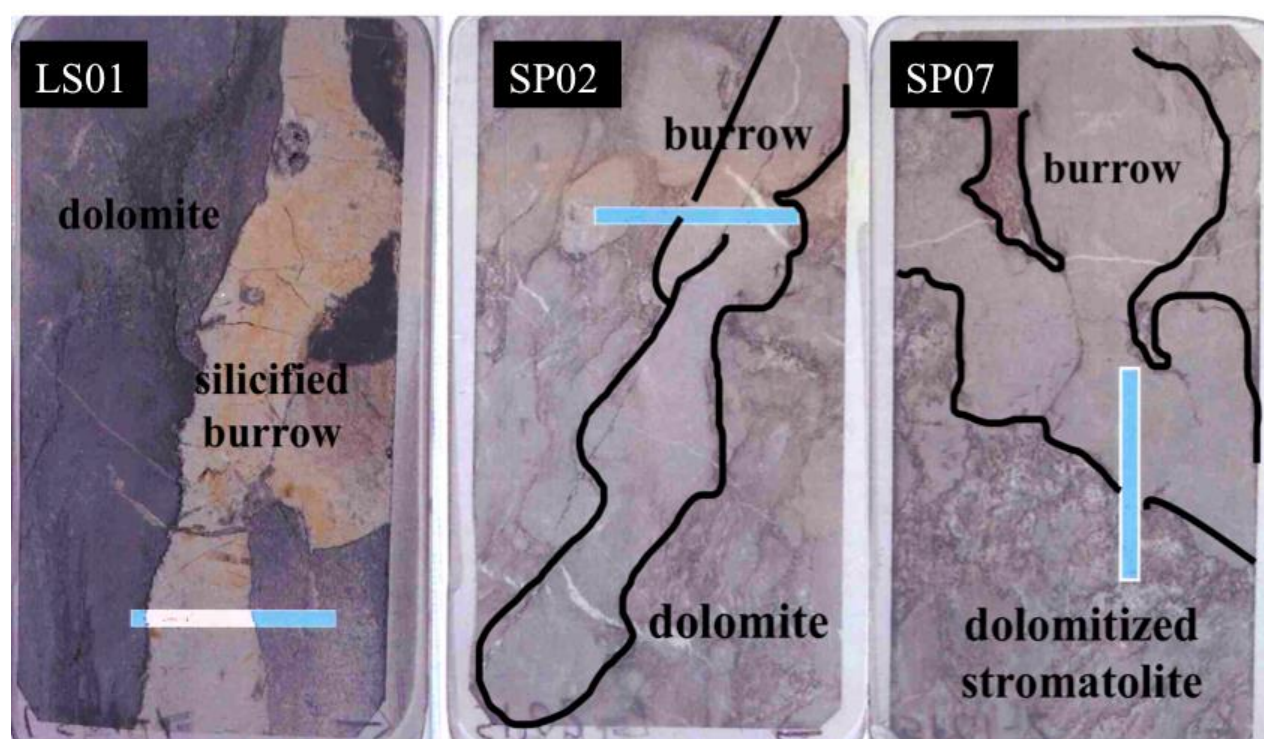
Figure A.2. Thin sections used for analysis in Chapter 2 for Figures 2.13, 2.14, and 2.15. LS01 is a thin section made from a polished slab of dolomite containing silicified burrows from the South Lakeside Mountains. The blue bar shows the section that was analyzed using QEMSCAN. SP02 is a thin section made from a polished slab of dolomite with dolomitized burrows from the southern House Range. SP07 is a thin section created from a polished slab of a dolomitized stromatolite containing dolomite burrows. Scale bars are 1cm.

⊟	Dolomite	98.80
	⊟ Dolomite	98.80
		Ca-Mg ox/CO3	96.20
		Ca-Mg ox/CO3(??)	2.50
		Ca-Mg(Fe) ox/CO3	0.05
		Mg-Ca-Fe ox/CO3	0.03
		Ca-Mg ox/CO3 ??	0.02
		Mg-Ca-Fe ox/CO3(??)	0.00
⊟	Amphibole	1.02
	⊟ Amphibole	1.02
		Ca Silicate ??	0.75
		Vermiculite?(1??)	0.11
		Al.K Silicate ??	0.10
		Al.Ca.K Silica ??	0.05
		Diopside	0.00
⊟	Background	0.11
	⊟ Background	0.11
		Background	0.11
		Si.Cl ??	0.00
⊟	Apatite	0.05
	⊟ Apatite	0.05
		Apatite	0.05
⊟	Calcite	0.04
	⊟ Calcite	0.04
		Ca ox/CO3	0.02
		Ca ox/CO3(??)	0.01
		Ca-Fe ox/CO3	0.01
⊟	Quartz	0.04
	⊟ Quartz	0.04
		Si ??	0.02
		SiH	0.01
		Quartz	0.01
		SiH(Fe)	0.00
⊟	Plagioclase	0.02
	⊟ Plagioclase	0.02
		Feldspar_Plag(??)	0.01
		Al-Silicate ??	0.01
⊟	Sulphates	0.01
	⊟ Sulphates	0.01
		Ca-S/SO4 ??	0.01
⊟	Fe-Ox/CO3	0.01
	⊟ Fe-Ox/CO3	0.01
		Fe Oxides(G07)	0.01
		Fe-oxides(??)	0.00

Figure A.3. This is the mineral percentage breakdown for sample SPO2 that is used in Chapter 2, Section 2.19, and Figure 2.17. This thin section was created from a burrowed and dolomitized stromatolite.

Figure A.4. This is the mineral percentage breakdown from sample SPO3, which was used in Chapter 2, Section 2.19, and is the sample used in Figure 2.18. This was created from a rock slab of dolomite containing dolomite burrows.

Quartz		51.75
	Quartz	51.75
	SiH	39.01
	Quartz	11.96
	Si ??	0.73
	SiH(Fe)	0.05
	SiH(KSn)	0.01
Dolomite		47.05
	Dolomite	47.05
	Ca-Mg ox/CO3	45.47
	Ca-Mg ox/CO3(??)	0.94
	Ca-Mg(Fe) ox/CO3	0.60
	Ca-Mg ox/CO3 ??	0.02
	Mg-Ca-Fe ox/CO3	0.01
	Mg-Ca-Fe ox/CO3(??)	0.01
Amphibole		0.63
	Amphibole	0.63
	Ca Silicate ??	0.39
	Vermiculite?(1??)	0.19
	Al.K Silicate ??	0.03
	Diopside	0.01
	Actinolite/Augite	0.00
	Wollastonite	0.00
	Hornblende3(CaD)	0.00
	Hornblende3	0.00
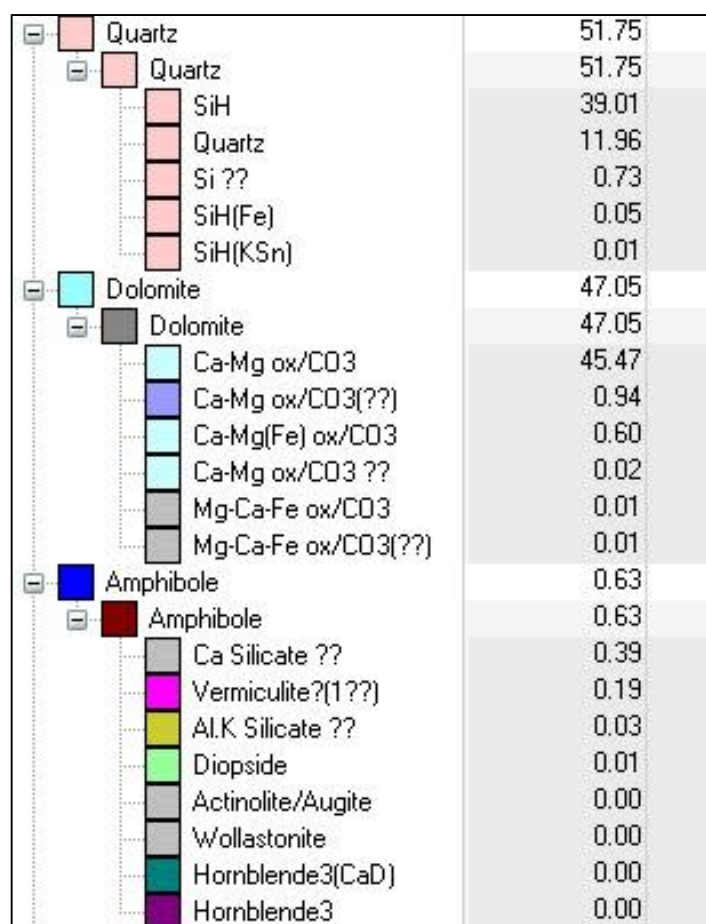
Figure A.5. Mineral percentage breakdown from sample LSO1, which was used in Chapter 2, Section 2.19, and is the sample used in Figure 2.16. This was created from a rock slab of dolomite containing silicified burrows.

Table A.1. The table shows the element distribution for all samples (LS01, SP02, SP07) and shows the percentage of each element found when all the sample percentage data is combined. These data are used in Chapter 2 in Section 2.19 to better understand which elements are the most prevalent in samples that were analyzed using QEMSCAN.

Size Fraction	Element	Sample %
-1000/+0	Al	0.05
	C	6.4
	Ca	10.84
	Fe	0.21
	Mg	6.47
	O	52.47
	P	0.03
	Si	23.43

Table A.2. Mineral distribution percentages for specified minerals when all the samples analyzed with QEMSCAN are combined.

	All	Background	Quartz	Olivine	Feldspar	Epidote
Area	33808	3.00	17496.00	6.00	1.00	5.00
Area Percent		0.01	51.75	0.02	0.00	0.01
Associations	1177.00	1442.00	790.00	23.00	4.00	18.00
Associations %	100.00	100.00	67.12	1.95	0.34	1.53
Background Transitions	110.00	0.00	4.00	0.00	0.00	0.00
Density	2.73	0.00	2.62	2.75	2.56	3.45
Host Associations	0.00	1430.00	0.00	0.00	0.00	0.00
Host Associations Percent	100.00	100.00	0.00	0.00	0.00	0.00
Intercepts	52.00	3.00	218.00	6.00	1.00	5.00
Mass	100.00	0.00	49.68	0.02	0.00	0.02
Mass Percent		0.00	49.68	0.02	0.00	0.02
Max Intercept	0.00	1.00	27.00	1.00	1.00	1.00
Min Intercept	0.00	1.00	1.00	1.00	1.00	1.00
Orientation	-4.32					
Perimeter Pixels	1424.83					
PSSA	0.31	200.00	2.49	200.00	200.00	200.00
Shape Factor	60.05					
Short Axis	2014.32					
Size	19506	30.00	2407.71	30.00	30.00	30.00
Surface Area	11.27	0.00	47.26	1.30	0.22	1.08
Surface Area %		0.00	3.64	0.00	0.00	0.00
Total Intercept	676220	60.00	349920.00	120.00	20.00	100.00
Volume	36.64	0.00	18.96	0.01	0.00	0.01
Volume %		0.01	51.75	0.02	0.00	0.01

Table A.3. Mineral distribution percentages for specified minerals when all the samples analyzed with QEMSCAN are combined.

	Silica	Calcite	FeO/CO3	Micas	S04	Apatite	Dolomite
Area	37.00	69.00	5.00	12.00	2.00	41.00	15906.00
Area Percent	0.11	0.20	0.01	0.04	0.01	0.12	47.05
Association	140.00	164.00	17.00	48.00	8.00	125.00	581.00
Association	11.89	13.93	1.44	4.08	0.68	10.62	49.36
Background Transitions	0.00	0.00	0.00	0.00	0.00	0.00	106.00
Density	2.70	2.71	5.15	3.09	2.30	3.19	2.84
Host Association	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Host Association Percent	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Intercepts	34.00	45.00	5.00	12.00	2.00	32.00	214.00
Mass	0.11	0.20	0.03	0.04	0.00	0.14	48.96
Mass Percent	0.11	0.20	0.03	0.04	0.00	0.14	48.96
Max Intercept	2.00	4.00	1.00	1.00	1.00	3.00	29.00
Mean Intercept Length	21.76	30.67	20.00	20.00	20.00	25.63	1486.54
Min Intercept	1.00	1.00	1.00	1.00	1.00	1.00	1.00
PSSA	183.78	130.43	200.00	200.00	200.00	156.10	2.69
Short Axis							
Surface Area	7.37	9.75	1.08	2.60	0.43	6.94	46.39
Surface Area %	0.00	0.00	0.00	0.00	0.00	0.00	96.36
Volume %	0.11	0.20	0.01	0.04	0.01	0.12	47.05

Table A.4. Mineral associations with Fe-Ox/CO3 that are used to determine how much iron is present in the minerals and the rock. These data are mostly used to evaluate stylolites and the dark colors representing them. These data were also used in the diagenetic analysis of the burrows verses the surrounding sediments in Chapter 2 in Sections 2.16 through Section 2.19.

Sub-group	Association	Percentage	Found in
Fe-Ox/CO^3	Dolomite	40%	Dolomite
	Quartz	30%	Quartz
	Micas	20%	Micas
	Others	10%	