PHOTOCOPY AND USE AUTHORIZATION

In presenting this thesis in partial fulfillment of the requirements for an advanced degree at Idaho State University, I agree that the Library shall make it freely available for inspection. I further state that permission for extensive copying of my thesis for scholarly purposes may be granted by the Dean of the Graduate School, Dean of my academic division, or by the University Librarian. It is understood that any copying or publication of this thesis for financial gain shall not be allowed without my written permission.

Signature ___________________________________

Date _______________________________________

LARGE SPONGES IN THE PHOSPHORIA FORMATION, WYOMING: THE PERMIAN GLASS FACTORY OF WESTERN LAURENTIA

By

William C. Moynihan

A thesis

Submitted in partial fulfillment

of the requirements for the degree of

Master of Science in the Department of Geosciences

Idaho State University

Summer 2018

To the Graduate Faculty:

The members of the committee appointed to examine the thesis of William C. Moynihan find it satisfactory and recommend that it be accepted.

Leif Tapanila,
Major Advisor

Paul Link,
Committee Member

Zac Gershberg,
Graduate Faculty Representative

ACKNOWLEDGMENTS

I would first like to thank my advisor Leif Tapanila for his guidance and motivation throughout the duration of this project, through its initial development, fieldwork, presentations, and the writing process. It was Leif who first suggested that the cylindrical chert in northwest Wyoming is sponge body fossils, which turned into the main argument of my thesis. I credit Leif for inspiring this project, and this is hopefully the first of many projects aimed at understanding the complex Phosphoria Formation.

I would also like to thank my secondary advisor Paul Link for his invaluable guidance throughout this project. Paul joined me in the field last summer, and exposed me to a variety of resources that have aided in the completion of this thesis, and helped to deepen my understanding of western North American geology as a whole.

Financial support for this project was provided by the Idaho State University Geosciences Geslin Award, and the Tobacco Root Geological Society. Many thanks for making this project possible.

I would like to thank the entire faculty and staff of the Idaho State University Geosciences Department, for creating a fantastic community. I especially want to thank Kate and Bre in the office for helping me schedule travel, and helping to resolve a variety of issues that came up through this project.

Lastly, I would like to thank my parents for their continued support, and their generous aid to help financially support me through two unpaid summers of research.

TABLE OF CONTENTS

LIST OF FIGURES

LARGE SPONGES IN THE PHOSPHORIA FORMATION, WYOMING: THE PERMIAN GLASS FACTORY OF WESTERN LAURENTIA

Thesis Abstract – Idaho State University (2018)

Large densely-packed cylindrical chert concretions were analyzed in detail in the Rex and Tosi Chert Members of the Phosphoria Formation in the Gros Ventre and Teton Ranges, Wyoming. Analysis of outcrops, hand samples, and thin sections leads to a new interpretation of the cylindrical chert concretions as body fossils of siliceous sponges preserved in life position, rather than large burrows. These are the first sponge body fossils reported in large numbers in the Phosphoria Basin, and record the glass factory of an extensive glass ramp depositional system. In situ production of siliceous sponge spicules outpaced and buried carbonate producers, and accumulated to form the thick, spiculitic Rex and Tosi Chert Members in a shallow-water lower ramp environment. The thick bedded cherts of the Phosphoria Formation represent the southernmost extent of the Permian spiculite belt, one of the largest accumulations of chert in earth history, which records the Permian Chert Event.

Keywords: Phosphoria Formation, Permian, sponges, chert, spiculite, glass ramp, Permian spiculite belt, Permian Chert Event

INTRODUCTION

The middle-late Permian Phosphoria Formation is a world-class sedimentary phosphorite deposit located in the northern Rocky Mountains of Idaho, Wyoming, Montana, Utah, and Nevada (Fig. 1). The strata are composed mainly of phosphorite, organic-rich mudrock, fine sandstone, dolostone, and thick chert intervals. The formation comprises the western U.S. Phosphate Field, one of the largest mined phosphate resources in the world (Jasinski et al., 2004). Because of the interest in mining, the organic-rich sedimentary phosphorites have been studied in great detail for the last century (Blackwelder, 1916; McKelvey, 1946; Hein et al., 2004). The Phosphoria Formation was first extensively mapped and sampled by the United States Geological Survey (USGS) in an effort to better understand its accumulation and the aerial extent of phosphate deposits (Richards and Mansfield, 1912; Cheney et al., 1951; Peterson et al., 1953; Sheldon, 1963; Sheldon et al., 1963; Hein et al., 2004 and references therein). These resource-focused studies emphasized the phosphatic shale members of the formation that dominate the thickest Phosphoria exposures in the central parts of the basin. Fewer studies (Carroll et al., 1998; Murchey, 2004) have focused on the chert members, and the relatively phosphate-poor, carbonate-rich deposits that make up the majority of the stratigraphy in the thinner, proximal exposures of Wyoming.

Within the chert members of the Phosphoria Formation in northwest Wyoming, there are parts of the stratigraphy that contain abundant, large cylindrical chert nodules, oriented perpendicular or oblique to bedding (Figs. 10, 11, and 15). Sheldon (1963) was the first to describe what he called ‘tubular chert’, and Cressman and Swanson (1964) hypothesized that these large structures are burrows. This hypothesis has held to the present day, and these particular cylindrical structures were given the trace fossil name *Skolithos grandis* (Andersson and Sauvagnat, 1998). Other origins have been suggested, including from inorganic processes related to sediment dewatering and compaction (Gableman, 1955; Yochelson, 1968), and as body fossils of siliceous sponges (Gutschick and Suttner, cited in Bromley et al., 1975). New evidence presented here based on detailed observations in outcrop, hand sample, and thin section supports the interpretation that these cylindrical chert structures in the Phosphoria Formation are siliceous sponge body fossils, preserved in a variety of forms.

This study presents four logged stratigraphic sections from northwest Wyoming, in deposits near the Phosphoria Basin margin (Figs. 1, 2). The focus is on understanding the accumulation of the chert units within the Phosphoria Formation, and their global context in the middle to late Permian. Evidence from stratigraphic sections, hand samples, and thin sections has led to the development of a new depositional model for chert accumulation, and provided insight into the paleooceanographic conditions in the Permian Phosphoria sea.

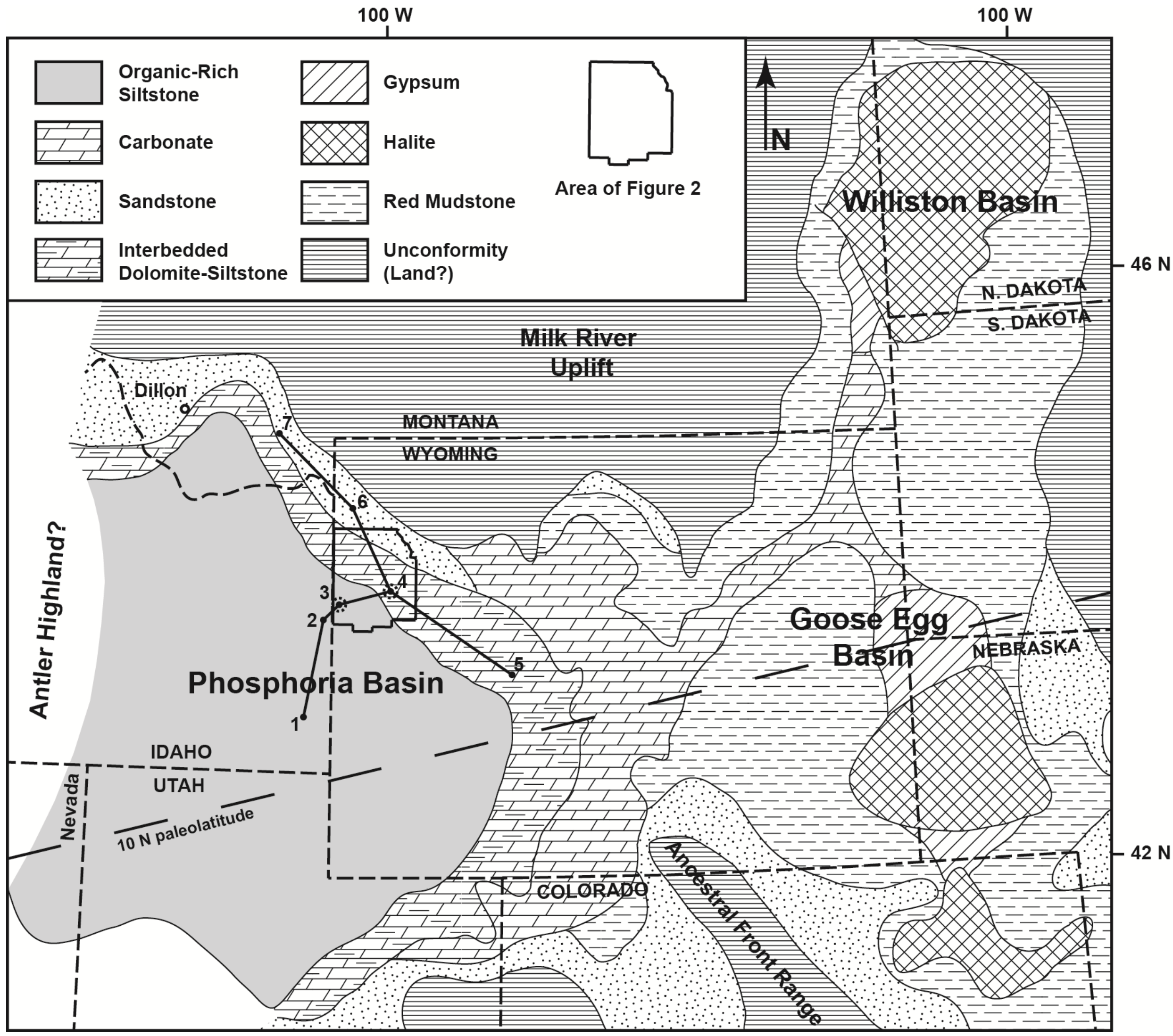


Figure 1. Map of the generalized distribution of Permian strata in the Phosphoria and Goose Egg basins. Modified from Carroll et al. (1998). Outline of Figure 2, and cross section line for Figure 3 are shown.

METHODS

The most complete exposures of the Phosphoria Formation were identified in the Gros Ventre and Teton Ranges in northwest Wyoming (Fig. 2) in summer 2017 based on previous studies, field guides, and geologic maps. Descriptions of outcrop locations are lacking in many previous studies, and the USGS studies often relied on excavated trenches, which are now mostly filled in. This study includes data from three of the most complete exposures in the northern

Gros Ventre Range, and one locality in the southern Teton Range, Teton County, Wyoming (Fig. 2).

Four stratigraphic sections (Fig. 4) were logged in the summer and fall, 2017 in order to better constrain the fossil assemblages and facies stacking patterns in the near-shore deposits of the Phosphoria Formation. Sections were logged with a Brunton compass, Jacob Staff, and a 10x hand lens. Bed thickness, lithology, sedimentary structures, grain size, shape and sorting, fossil content, and deformational structures were recorded. Hand samples of each facies were collected throughout the stratigraphy at three localities (Flat Creek, Gros Ventre Slide, and Ski Lake) (Fig. 2). A U.S. Forest Service Permit restricted sample collection to non-wilderness areas, so no samples were collected at the best and most extensive exposure, Crystal Creek. Thin sections were made of each sample in order to identify microfacies and microfossil assemblages. Many hand samples of cylindrical chert nodules were cut and polished, and analyzed under stereoscope in order to reveal internal features in greater detail.

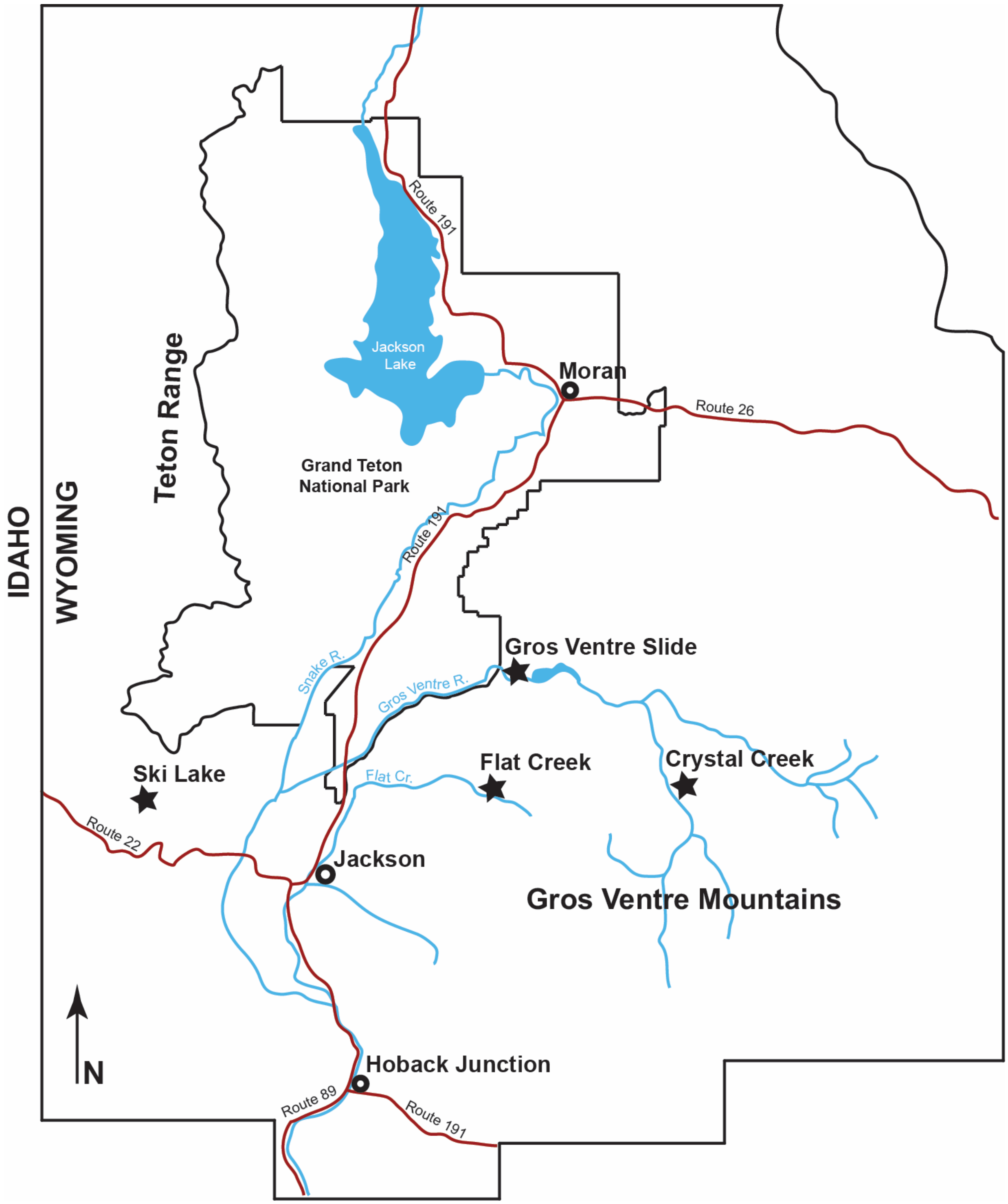


Figure 2. Map of the southern part of Teton County, Wyoming. Outcrops logged for this study are shown by stars and labeled.

REGIONAL GEOLOGY

Tectonic Setting

The Phosphoria Basin covered an area about 400,000 km^2 on the western margin of North America (Maughan, 1984), and was bounded to the north by the Milk River uplift in Montana, to the east by extensive evaporative basins (Goose Egg basin), and to the south by the Front Range uplift of the Ancestral Rockies (Fig. 1). The basin is thought to have been bounded on the west by a poorly defined island remnant of the Antler Highland (Wardlaw and Collinson, 1986; Piper and Link, 2002), and Phosphoria strata are missing west of central Idaho (Fig. 1). In the Pennsylvanian to early Permian, the Wyoming shelf was dominated by eolian sand dunes (Stephens and Carroll, 1999), prior to Phosphoria deposition. Beginning in late Cisuralian (Kungurian) time (Fig. 3), a marine transgression flooded the Pennsylvanian-early Permian Wood River, Sublett and Oquirrh basins, and much of the continental shelf in Wyoming, forming the epicratonic Phosphoria Basin (Stephens and Carroll, 1999). The Phosphoria Formation strata cover about 350,000 km^2 (Hein et al., 2004) within the basin, and were deposited between the late Kungurian (Cisuralian) and the early Wuchiapingian (Lopignian) ages in the Phosphoria sea (Fig. 3) (Wardlaw, 2015; Fig. 3 in Davydov et al., 2016). Continuous deposition within the basin persisted for most of the middle-late Permian, and possibly until the end-Permian extinction (Davydov et al., 2016). Time constraints are discussed in greater detail in a later section.

The Phosphoria Formation thickens from east to west, and was deposited on an eastern inner shelf and a deeper-water western shelf slope and basin (Davydov et al., 2016). The Phosphoria Basin was located in the equatorial zone, between about 10° and 20° N paleolatitude (Stephens and Carroll, 1999) allowing for a carbonate ramp to develop on its eastern and southern margin (Fig. 1) (Wardlaw and Collinson, 1986). A lack of tidal channeling and sedimentary structures suggests that the ramp was a widespread area of low-energy deposition (Wardlaw and Collinson, 1986). According to Ahr (1973), the carbonate ramp gradually sloped westward at an angle of less than 1°, and no significant break in slope is recognized. Furthermore, regional stratigraphic relationships suggest that the ramp in Wyoming was much more shallowly west-dipping at 11-12 cm/km (0.01°) (Stephens and Carroll, 1999). Palinspastic reconstructions of the Sevier thrust belt near the Idaho-Wyoming border (Peterson, 1984; Royse, 1993) suggest a gradually sloping ramp, and there is no sedimentological evidence of a significant break in slope.

The widespread occurrence of particular marker beds, like the basal ‘fish scale’ bed for example, has also been used to argue for a shallowly-sloping ramp. Furthermore, Whalen (1996) interpreted facies patterns to indicate a homoclinal ramp in the north, grading into a low-gradient distally-steepened ramp to the south in Utah, both of which deepen toward the west. In this study, the Phosphoria Formation is interpreted to have been deposited on a shallowly west-sloping homoclinal ramp (Ahr, 1973).

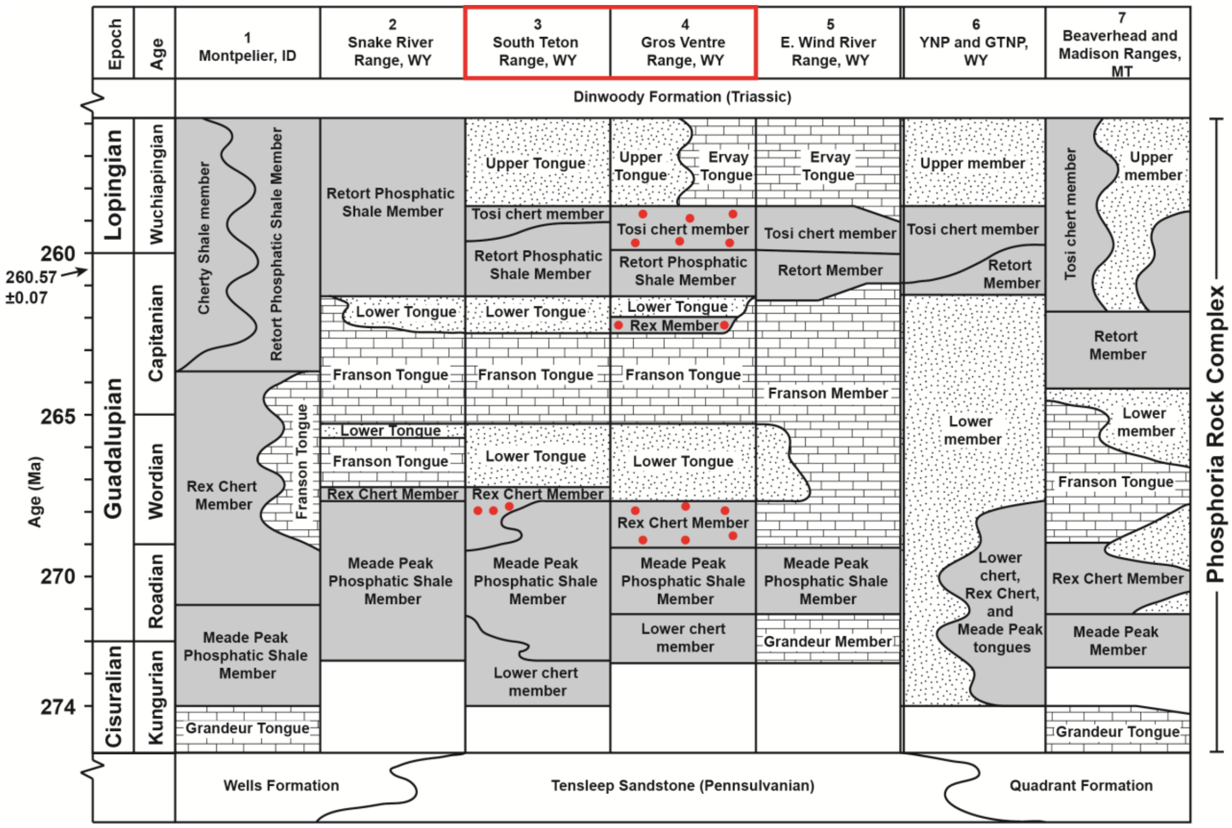


Figure 3. Stratigraphic correlation diagram following the cross section lines in Figure 1. Phosphoria Formation is in gray, limestone is the Park City Formation, and sandstone is the Shedhorn Formation. The two columns in northwest Wyoming are highlighted in red. Red dots indicate where cylindrical chert occurs. Modified from Cressman and Swanson (1964) and Sheldon (1963).

Lithostratigraphy

The name Phosphoria Formation is lithostratigraphically defined, and is reserved only for phosphorite, dark mudstone, and chert (Cressman and Swanson, 1964), though other lithologies commonly interfinger with these in Montana, Wyoming, and Utah. The limestone and dolomite facies comprise the Park City Formation, and the Shedhorn Sandstone is defined as the sandstone units present in the northern part of the basin (Cressman and Swanson, 1964). These formations are all subdivided into members, which extend only as far as the units maintain their lithologic character. These formation names can be confusing, given the complex interfingering relationship of quite different facies (Fig. 3), especially when correlating the stratigraphy between the middle of the basin and its margins. Furthermore, few beds within the stratigraphy fall into one of these end-member lithologies, and mixed carbonate-siliciclastic deposits are common. Some studies have grouped all Phosphoria Basin deposits under the name Phosphoria Rock Complex (e.g. Yochelson, 1968; Pommer, 2017). Because the sections measured in this study contain mostly Phosphoria strata, and to remain consistent with most previous studies, all of the stratigraphy in the logged sections in this study will be referred to as the Phosphoria Formation, unless otherwise specified. Individual member names are assigned according to criteria outlined in previous studies.

In southeast Idaho, the Phosphoria Formation overlies the Pennsylvanian Wells Formation, and in Wyoming, it overlies the Pennsylvanian Tensleep Formation. Further north in Montana, it overlies the Quadrant Formation in certain areas (Fig. 3). Following deposition, it is thought that late Permian to Early Triassic erosion variably removed the upper part of the Phosphoria strata (Wardlaw and Collinson, 1986). Following this hiatus, the Dinwoody Formation was deposited beginning in the early Triassic, overlying the Phosphoria Formation for most of its aerial extent (Fig. 3).

The major units (members) of the Phosphoria Formation in stratigraphic order are: the Lower Chert, Meade Peak Phosphatic Shale, Rex Chert, Retort Phosphatic Shale, and Tosi Chert Members (McKelvey et al., 1959; Yochelson, 1968). Sheldon (1963) divided the formation into two units, lower (Lower Chert, Meade Peak, and lower part of the Rex Member) and upper (upper part of the Rex, Retort, and Tosi Members), stratigraphically separated by tongues of the Park City and Shedhorn Formations (Fig. 3). In some western exposures in central Idaho, the Meade Peak Phosphatic Shale Member is locally overlain only by a single thick informal unit called the Cherty Shale member (McKelvey et al., 1959). The members vary greatly in lithic character throughout the basin, and some studies even correlate two different members, like the Retort and Meade Peak Members, across the basin (Wardlaw and Collinson, 1986; Davydov et al., 2016). For this study, the members defined by McKelvey et al. (1959) will be used and are described in detail below, as observed in northwest Wyoming.

Lower Chert

The Lower Chert Member is defined as a resistant, dominantly cherty and slightly phosphatic unit that overlies the Tensleep Formation and it is overlain by the non-resistant black shale of the Meade Peak Member (McKelvey et al., 1959). The unit consists of gray chert and dolostone, contains abundant siliceous sponge spicules in some beds, and has a maximum thickness of about 12 m in the southern Wyoming Range, south of the present study area. It is confined to a narrow N-S-trending belt in western Wyoming, is only about 3 m thick in the southern Teton Range, and pinches out in the northern Gros Ventre Range between Flat Creek and Crystal Creek (McKelvey et al., 1959). To the east the Lower Chert Member grades into the upper part of the Grandeur Member of the Park City Formation, and to the west, it grades into the lower part of the Meade Peak Member (Sheldon, 1963). Sheldon (1963) suggested that this is a transgressive unit, and is younger to the east.

Meade Peak Phosphatic Shale

The Meade Peak Phosphatic Shale Member is the most aerially extensive unit of the Phosphoria Formation (Wardlaw and Collinson, 1986), and it is composed of dark carbonate, phosphatic, and argillaceous rocks, with end-member lithologies of black mudstone and phosphorite (McKelvey et al., 1959). However, the sequence of rock types, the facies, and the thickness are diverse throughout the basin (Sheldon, 1963). The Meade Peak Member is between 38 m and 69 m thick in southeast Idaho, and thickens southward to over 90 m in the Wasatch Range of northern Utah. The member thins to the west to 26 m in the Cassia Mountains, and thins to the north and east of southeast Idaho before pinching out in southwest Montana and western Wyoming. It is no more than 6 m thick in the Gros Ventre Range (Sheldon, 1963). Within the Meade Peak Member, finely pelletal phosphorite is most abundant in southeast Idaho, but oolitic, pisolitic, and bioclastic phosphorite are more common toward the basin margins (McKelvey et al., 1959). The ore zones and key beds within the Meade Peak can be correlated over much of southeastern Idaho, though composition and thickness vary locally. Toward the east in central Wyoming, the upper and lower parts of the formation pass into chert (Lower Chert and Rex Chert) and carbonate (Park City Formation), leaving a thinner phosphatic middle section (Fig. 3) (McKelvey et al., 1959). The unit also is sandier to the north and east, and bioclastic apatite grains become more prominent (Sheldon, 1963). The lowermost bed for most exposures is a thin phosphorite known as the ‘fish scale’ bed, containing abundant inarticulate brachiopods, phosphatic fish scales, bones, and nodules (McKelvey et al., 1959).

In the type area in southeast Idaho, the vertical sequence of facies within the Meade Peak is symmetrical with a phosphate-poor, interpreted deeper-water middle zone (Sheldon, 1957; 1963), known in the mining jargon as the ‘center waste shale’. From southeast Idaho, the transgressive base of the Meade Peak is slightly younger moving to the south and to the east into Wyoming (Wardlaw and Collinson, 1986).

Rex Chert

The Rex Chert Member is defined by McKelvey et al. (1959) as the thick interval of dark resistant chert beds that overlies the Meade Peak Member. The chert in this unit is thin-bedded to massive, varies from black to nearly white in color, and contains various impurities such as carbonate, quartz sand, calcite, dolomite, and apatite. About 20% of chert beds contain megascopic sponge spicules (Sheldon, 1963).

In the shallow ramp area of southeast Idaho, the Rex Chert is about 49 m thick, and thins to the north, east, and south, where it passes into carbonate rock of the Franson Member of the Park City Formation (McKelvey et al., 1959). To the west, the Rex Member thickens drastically in the Sublett Range. In the Wyoming and Gros Ventre Ranges in western Wyoming, it is about 6 to 12 m thick, but thins to only a few meters in the southern Teton Range (Sheldon, 1963). The upper and lower contacts of the Rex are hard resistant chert, and are older to the north and east of its type section in southeast Idaho (McKelvey et al., 1959). Because of this, the Rex Chert has been interpreted as a regressive unit. The field area in northwest Wyoming likely contains mostly Franson Member strata in the middle of the exposed stratigraphic section (Fig. 3), but for simplicity, it will still be referred to as the Rex Chert Member in this study, unless otherwise specified.

Retort Phosphatic Shale

The Retort Phosphatic Shale Member is defined simply as the phosphatic shale unit in the upper part of the formation (McKelvey et al., 1959). The lithic character throughout the basin is almost identical to the Meade Peak Member (Sheldon, 1963). The strata can be divided into a lower phosphatic zone, a thinner middle slightly calcareous mudstone, and an upper phosphatic zone (McKelvey et al., 1959). The phosphorite is dominantly pelletal, but also commonly contains bioclastic apatite grains (Sheldon, 1963).

At its type locality in southwest Montana, the Retort Member is 18 m thick, and consists of thin-bedded dark carbonaceous mudstone and pelletal phosphorite (McKelvey et al., 1959). In the Gros Ventre Mountains the Retort is generally 11 m thick or less, and thins to the north, but thickens further east in the Wind River Range (McKelvey et al., 1959). The variation in thickness is largely due to changes in thickness of the mudstone, which is thickest in the Gros Ventre and Wind River Ranges. Phosphorite beds are less variable in thickness throughout the extent of the Retort Member (Sheldon, 1963).

The upper contact is often gradational with the overlying Tosi Member in this area, and the lower contact with the Rex Member often appears conformable as well (McKelvey et al., 1959). The Retort Member is thought to represent widespread transgression throughout the basin (Wardlaw, 1979). Deposition was focused in the northern and eastern parts of the basin, and strata are absent to the south in Utah.

Tosi Chert

The Tosi Chert Member is defined as the thin chert beds that overlie the Retort Member, and is present in Montana and Wyoming, and in limited areas of eastern Idaho and northern Utah. At its type locality in the southern Gros Ventre Range, the Tosi Chert consists of a thick lower grayish-brown to brownish-black chert interval overlain by a thinner light-gray sandy chert interval (McKelvey et al., 1959).

The Tosi Chert Member is 10 m thick at its type locality in the southern Gros Ventre Range, and overlies the Retort Member (McKelvey et al., 1959). This contact is often gradational, and consists of interbedded fissile mudstone and thin-bedded black chert, grading up into entirely bedded black chert. To the southwest, the Tosi Chert grades into dark gray mudstone and is included in the Retort Member. To the north and east, the Tosi thickens to as much as 44 m in Montana, and the bottom gradational contact moves lower in the section. In the Gros Ventre and Teton Ranges, the Upper Member of the Shedhorn Sandstone overlies the Tosi Member, and is overlain by the base of the Dinwoody Formation (Fig. 3).

Biostratigraphy and Chronostratigraphy

The only age constraints on the Phosphoria Formation deposition come from conodonts (Wardlaw and Collinson, 1986; Wardlaw, 2015), and more recently, from a single volcanic ash bed within the upper Meade Peak Member (Davydov et al., 2016). Wardlaw and Collinson (1986) identified 8 biostratigraphic zones based on brachiopod and conodont fossils in the Phosphoria and interfingering Park City Formation. After further refinement of these biostratigraphic zones, and better age-constraints on conodont faunas, Wardlaw (2015) bracketed the deposition of the Phosphoria Formation to 8.9 m.y. between 274.0 Ma and 265.1 Ma. Davydov et al. (2016) dated a volcanic ash bed in the upper Meade Peak Member in the Cassia Mountains, southern Idaho, with a U/Pb age of 260.57 ±0.07 Ma. This greatly extends the depositional age of the Phosphoria Formation, and suggests that it may span over 20 m.y. until the end Permian. An added complexity is that the Meade Peak Member in the Cassia Mountains is correlative with the Retort Member in Wyoming (Wardlaw and Collinson, 1986; Davydov et al., 2016), and the lithostratigraphically defined members of the formation yield complex lateral relations of units. Pommer (2017) obtained U/Pb dates from carbonate material of a brachiopod and micrite within the upper part of the formation, which agree with Wardlaw (2015), that Phosphoria deposition ceased at the end of the Wordian age (265 Ma).

In this study, Phosphoria deposition is interpreted to have spanned 14 m.y., starting at 274.0 Ma, and the Retort-Tosi boundary is about 260 Ma (Fig. 3). The overlying Tosi Chert is 12 m thick, where measured in this study. According to Hein et al. (2004), the sedimentation rate was unchanged for the duration of Phosphoria deposition, and was about 1 m/400-480 ky. This is certainly an oversimplification, but assuming the slowest sedimentation rate (1 m/480 ky) for the relatively thin deposits in northwest Wyoming, the Tosi chert would have been deposited over 5.76 m.y., ending at 254.24 Ma, still short of the end-Permian. Age constraints in the Phosphoria Formation remain poor, however, and new methods are needed to better constrain ages in these rocks.

Sequence Stratigraphy

The Phorphoria deposition spans one of the most drastic global climate shifts in earth history. During the Pennsylvanian and early Permian, there was widespread glaciation in the southern hemisphere, low atmospheric CO_2 levels, and likely sea ice in the northern hemisphere (Hiatt and Budd, 2001). The late Permian is marked by the onset of hot, dry climates for much of the middle latitudes, increased atmospheric CO_2 levels, widespread desert environments and evaporite deposition, and widespread anoxic conditions in the world's oceans (Hiatt and Budd, 2001). Sheldon (1957) suggested that this climate shift is responsible for cyclicity in eustatic sea level changes, and the resultant cyclicity of the Phosphoria strata.

The entire Phosphoria Formation deposition occurred within a second-order transgressive-regressive sequence over about 8-14 m.y. (Pommer and Sarg, 2017), flooding the

Wyoming shelf. Superimposed are three third-order sequences (Grandeur, Franson, Ervay), each about 2-5 m.y. duration (Whalen, 1996; Pommer and Sarg, 2017). Most previous studies recognized two transgressive-regressive sequences, represented by occurrence of the Meade Peak and Retort Phosphatic Shale Members on the Wyoming shelf (Sheldon, 1963). The transgression and deposition of the two thick shale members alternated with the progradation of the carbonate ramp out into the basin. Based on thicknesses between stratal surfaces and key beds, the sea-level is thought to have fluctuated by around 80 m in the Franson Member cycle and 45 m in the Ervay Member cycle (Pommer, 2018, personal communication). On a very shallowly-sloping carbonate ramp, this is magnitude of sea-level oscillation would result in large lateral migrations of lithologies during transgression and regression, and the resultant facies are highly variable in vertical sequence.

Phosphorite deposition largely dominated lowstands and transgressions (Pommer, 2017). Siliceous sponge spiculitic chert and bioclastic limestone were deposited during late transgressions and early highstands. During highstands, peritidal microbial communities, mollusks, and bioturbated muds and sands dominated the central Wyoming deposits, inland of spiculitic chert (Pommer, 2017).

Higher-order sequence stratigraphy is still poorly constrained in the Phosphoria Basin deposits because of the highly variable lithologies, and complex interfingering relationship of units. A general lack of peritidal facies and shallow-water sedimentary structures makes fine-scale sequence stratigraphy challenging as well.

Climate and Provenance

The Phosphoria Basin was largely starved of siliciclastic sediment, except for in northwest Wyoming and southwest Montana, where sand was sourced from the Milk River Uplift (Fig. 1) and deposited as the Shedhorn Sandstone (Wardlaw and Collinson, 1986). Carbonate and phosphatic sediment was produced in situ, and the relatively small portion of siliciclastic sediment is eolian dust, distributed throughout the basin.

The organic rich siltstones (shales) contain very little clay, and have grain size distributions remarkably similar to those in modern near-shore eolian-derived sediment off the coast of northwest Africa (Carroll et al., 1998). The ratios of major element oxides and trace elements in the shales are also similar to the world shale average, suggesting a solely terrigenous source for the siliciclastic fraction of sediment (Piper and Link, 2002). The clastic units lack sedimentary structures indicative of debris flows, channels, or even wave action, so settling of suspended material through the water column is the most likely mode of deposition (Carroll et al., 1998). Furthermore, a mean sediment accumulation rate for the terrigenous fraction of the Meade Peak Member is within the range of accumulation rates for modern pelagic sedimentary environments (Carroll et al., 1998). These observations all point to eolian sediment transport for the siliclastic fraction of sediment, which is consistent with an arid climate. The Phosphoria Basin was deposited in the tropics, between about 5° and 25° N paleolatitude (Fig. 1) (Piper and Link, 2002), analogous to arid modern North Africa.

There is an empirical relationship between arid paleoclimate and chert deposits, which is also consistent with observations of the Phosphoria Formation. To the east, the evaporative Goose Egg and Williston Basins correlate with Phosphoria deposition (Carroll et al., 1998). Sand dunes are not recognized during the inundation of the Wyoming shelf, but they were extensive prior to initial transgression and Phosphoria deposition (Stephens and Carroll, 1999).

The thickest shale (‘organic siltstone’) intervals are in the central part of the basin in southeast Idaho, suggesting that eolian sediment was largely bypassing the shallower-water carbonate deposits in Wyoming. If the eolian sediment flux was as high in the shallow Wyoming shelf deposits as it is in the deeper basin in Idaho, then either carbonate production would have been hindered, or the resultant deposits would be much thicker than in Idaho. There is fine sand and silt present in every facies within the formation, but the finer fraction is in higher concentrations in the central part of the basin in Idaho, represented by the thick phosphatic shale members.

Global circulation models from Early Permian and Triassic times suggest northerly or northeasterly wind directions in northwest Pangaea, pointing to a northerly source for sediment (Carroll et al., 1998). These wind directions agree with eolian paleowind indicators in Permian sandstones on the Colorado Plateau that predate Phosphoria deposition, so this circulation pattern is thought to be long-standing. Therefore, Carroll et al. (1998) suggest that sediment is sourced from the Milk River Uplift north and east of the basin in Montana. Furthermore, the lower part of the Shedhorn Sandstone, which fringes much of the northern basin margin (Fig. 1) has been correlated with the Meade Peak Member (Thornburg, 1990), and the compositional and textural maturity of the Shedhorn suggests possible reworking of eolian sands. Therefore, the Shedhorn Sandstone is the shore-face equivalent of the Meade Peak Member. There are Permian alluvial sandstones on the southern basin margin near the Colorado-Wyoming border, which were probably sourced from the ancestral Uncompahgre uplift in Colorado (Carroll et al., 1998).

Upwelling and Phosphogenesis

Phosphate ore is in the mineral form carbonate fluorapatite, and occurs most commonly as spheroids, and less commonly as fossils, bioclasts, crystals, and cement (Sheldon et al., 1963). Most studies agree that phosphorite was organically precipitated, and it was highly concentrated by plankton in the water column (Bentor, 1980; Sheldon, 1980). Phosphogenesis is the precipitation of phosphate from seawater at or just below the sediment-water interface, and has commonly been interpreted to represent a lower ramp, deep water environment (McKelvey et al., 1959; Wardlaw and Collinson, 1986), where phosphorites are thought to record periods of sediment starvation. Other studies however, have provided evidence for phosphogenesis throughout the entire basin, from deep basinal through subtidal and peritidal settings, and phosphorite is found all throughout the basin (Sheldon, 1984; Peterson, 1984; Stephens and Carroll, 1999; Hiatt and Budd, 2001; Hein et al., 2004; Pommer, 2017).

The nutrient-rich waters required for planktonic phosphate deposition, along with the presence of a cold-water brachiopod facies, suggest that the basin was a site of marine upwelling (Wardlaw and Collinson, 1986). Marine upwelling has been long accepted as a major driver of phosphate production, as cold nutrient-rich bottom waters are brought into the shallow basin, resulting in algal blooms, followed by widespread anoxia within the basin, and accumulation of phosphate-rich organic matter on the basin floor. Paleoclimate reconstructions support this model as well, as northerly surface winds (Carroll et al., 1998) have been interpreted to have a component of Ekman transport (away from shore) of surface waters in the basin, driving upwelling currents (Piper and Link, 2002). In the deeper basin, bacterial breakdown of shallowly buried organic matter released phosphorous into the sediment (Hiatt and Budd, 2001). Phosphorites in the shallow basin, however, record a significant influx of phosphorous into the shallow waters. Therefore, upwelling currents must have reached the shallow parts of the ramp (Hiatt and Budd, 2001). However, maximum upwelling took place on the middle ramp, and the release of phosphate during the breakdown of organic matter could have fed the productivity in the shallow waters. Upwelling currents are considered the main driver of phosphorite deposition basinwide (Piper and Link, 2002; Hiatt and Budd, 2001; Hein et al., 2004).

Biofacies

Although they are largely overlooked, siliceous sponge spicules are by far the most common allochems in the Phosphoria Formation, they occur in all members, and their total volume exceeds that of any other fossil group (Murchey, 2004). Strangely though, no sponge body fossils have been found within Phosphoria sections. The only sponges reported within the basin come from the Park City Formation, and are very rare and small (Yochelson, 1968; Rigby and Boyd, 2004).

In carbonate facies, inarticulate brachiopods and mollusks are the most abundant fossils, and bryozoans and cephalopods are less abundant, though still common (Yochelson, 1968). These groups however, are characterized by very low taxonomic diversity. Brachiopod faunas associated with the carbonate ramp in Wyoming are warm-water faunas and show close affinity to the Permian Basin faunas in western Texas (Wardlaw, 1977). Brachiopod faunas in limestones associated with deeper ramp facies toward the middle of the basin are closer in affinity to arctic faunas, and indicate cooler-water deposition, suggesting upwelling (Wardlaw and Collinson, 1986).

Bryozoans are abundant in most limestones, but once again, are low in diversity (Yochelson, 1968). Crinoid debris is limited, and rarely articulated (Wardlaw and Collinson, 1986). Disarticulated fish bones, scales, and most often teeth, are common throughout the formation (Yochelson, 1968; Kulas and Batten, 1997), and burrows are also common (Wardlaw and Collinson, 1986). Large cephalopods are rare, but occur in the Meade Peak Member. Large predatory *Helicoprion* sharks are found in the deeper-basin deposits of the Meade Peak Member in Idaho, and likely fed on cephalopods and the abundant fish (Tapanila et al., 2013). Corals are rare, and trilobites and fusulinids appear to be absent (Yochelson, 1968).

The fauna is seemingly unbalanced compared with slightly older Permian strata of the mid-continent, favoring brachiopods and mollusks, and the limited diversity is in striking contrast to diverse faunas in the Permian of western Texas (Yochelson, 1968). The fossil assemblages in the Meade Peak and Retort Members are strikingly similar, and only differ in degree of abundance, rather than kind of groups. In modern faunas, a lack of diversity, but abundance of individuals has been interpreted to mean less than ideal conditions for life (Yochelson, 1968). However, the abundance of fish remains and large predatory sharks at higher trophic levels, and the thick, organic-rich phosphorites suggest that the basin was home to a large biomass throughout the middle-late Permian.

Chert: Opal-A, Opal-CT, Quartz

The thick chert deposits in the Phosphoria Formation are thought to be formed from the accumulation of siliceous sponge spicules (Sheldon, 1963; Murchey, 2004). Siliceous sponges are silica biomineralizers, and they make their skeletons out of needle-like spicules made of hydrous, amorphous silica (opal-A). Spiculite is used herein to describe rocks composed primarily of siliceous sponge spicules. Spiculitic chert deposits that we observe in the rock record are made of fine crystalline quartz, so a phase change takes place after burial. This transition from opal-A to quartz has been well documented, and includes a middle, transitional cristobalite-trydamite (opal-CT) phase. Opal-A is hydrous and easily dissolved, opal-CT is microcrystalline, and quartz is fully crystalline and insoluble.

Chert has most often been attributed to deep-water basinal accumulation and recrystallization of siliceous oozes (Gates et al., 2004), but this model does not fit observations of many chert deposits, the Phosphoria Formation in particular. Ghosts of carbonate precursors in chert indicate that nodular chert forms by the replacement of carbonate or other sediment (Maliva and Siever, 1989). Furthermore, the dissolution of carbonate and precipitation of silica occur simultaneously along a thin solution film, and happen at the same rate. Although nodular chert occurs in limestones of a wide variety of environments, a correlation is seen in some formations between nodular chert and abundance of siliceous sponge spicules (Maliva and Siever, 1989). Therefore, a high concentration of spicule-bearing nodular chert likely represents an interval when sponges were living in that area.

Permian Chert Event

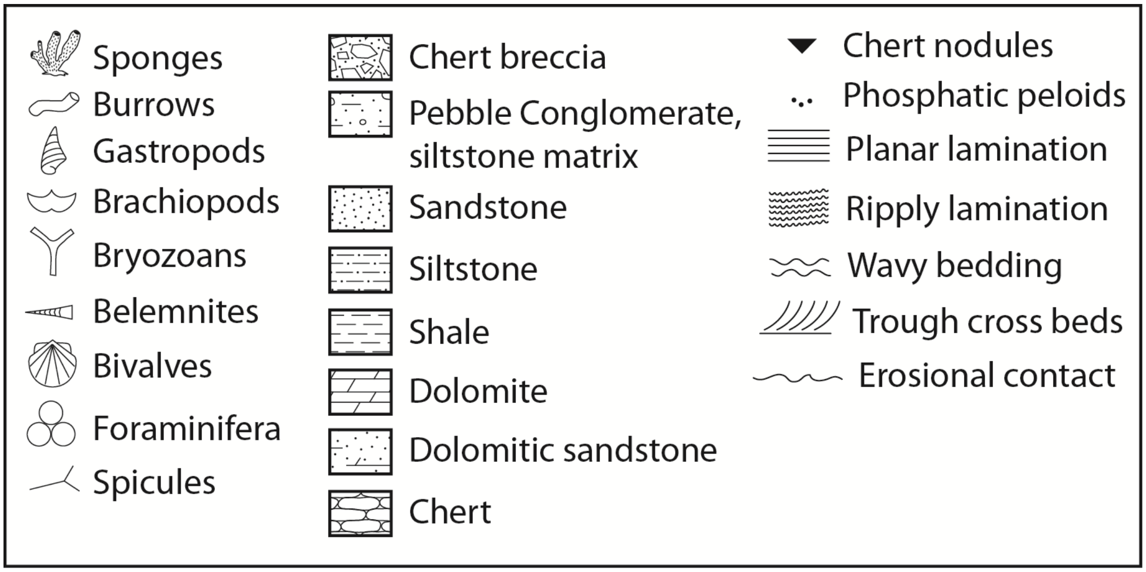
The chert members of the Phosphoria Formation are much less well-studied than the economic phosphatic shale members, yet chert or chert-bearing units make up roughly half of the stratigraphic thickness in western Wyoming. Chert is common in all parts of the basin, and it is thought to be formed both by primary deposition of biosiliceous sediment (black bedded chert), but mostly as varying degrees of replacement of carbonate (Wardlaw and Collinson, 1986). The facies distribution of chert deposits provides a potential record of the paleoenvironmental distributions of siliceous organisms (Maliva et al., 1989). It has been long recognized that the source of silica for the chert is mainly from siliceous sponge spicules (Sheldon, 1963;

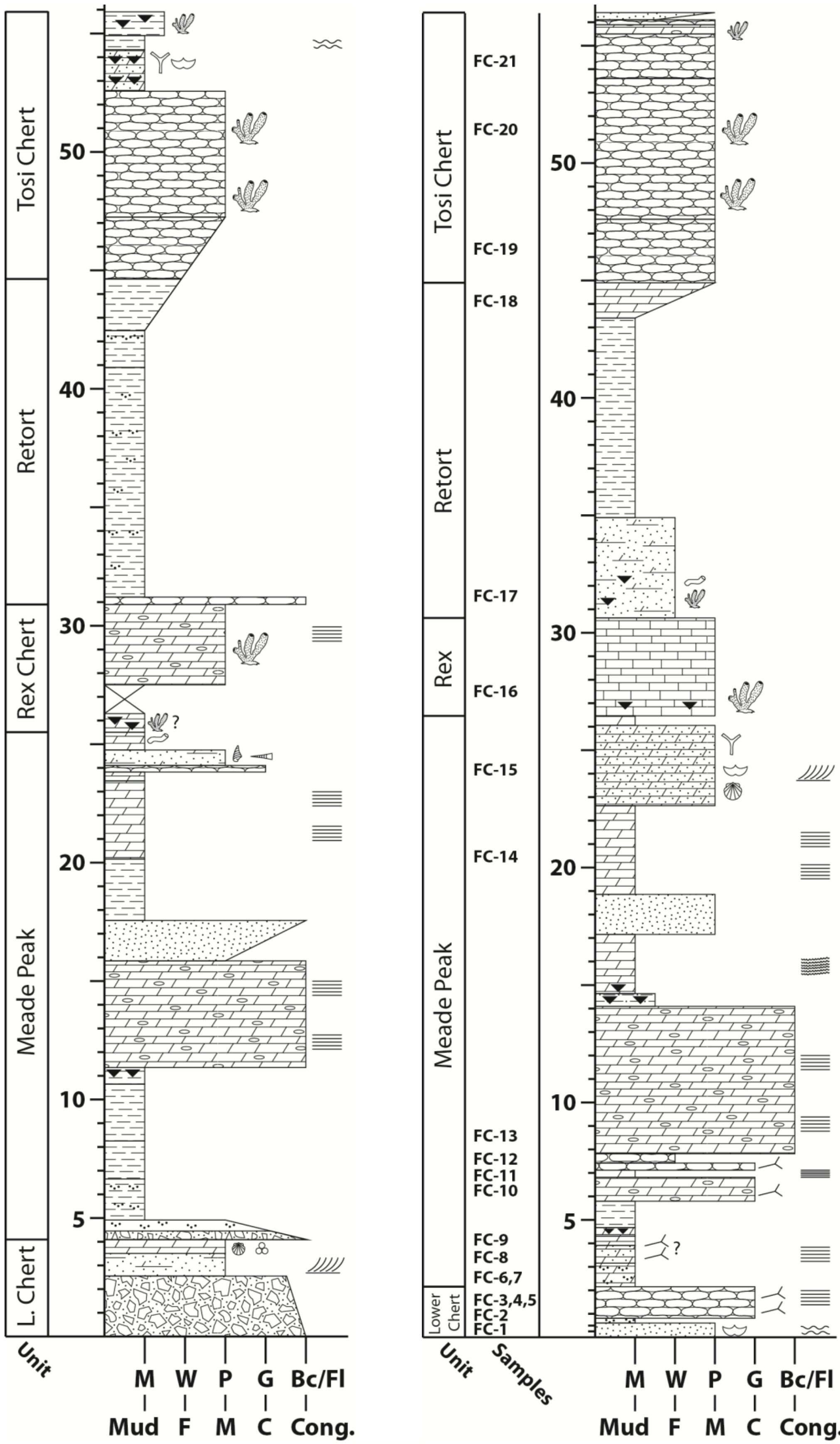
Yochelson, 1968), which are the principal allochems in the Rex and Tosi Chert Members, and common components throughout the formation (Murchey, 2004). However, no siliceous sponge body fossils have been reported in the Phosphoria Formation.

Rigby and Boyd (2004) described small twig-like sponges in the Phosphoria-equivalent Franson Member of the Park City Formation in Wyoming, but they are not present in large enough volumes to be responsible for the widespread spicules, and thick chert units. It is well established that sponges were present in the basin in large volumes, based on abundance of spicules (Murchey, 2004), and therefore it is likely that body fossils are preserved somewhere in the stratigraphy.

The middle to late Permian Chert Event (Murchey and Jones, 1992) was a time of widespread chert deposition extending far beyond the Phosphoria Basin. The thick spiculites extend from central Nevada through Idaho, Wyoming, Montana, Alberta, British Columbia, all the way to arctic Canada, and Svalbard, Norway (Fig. 20) (Murchey, 2004). These rocks collectively compose the Permian spiculite belt, and consist both of deposits on the Permian western margin of North America, and in terranes of the North American Cordillera (Murchey and Jones, 1992). Biogenic siliceous sediments within the spiculite belt were deposited in settings ranging from shallow drowned platforms to deep oceanic seamounts (Murchey and Jones, 1992). Radiolarians are not present in the Phosphoria Formation, and therefore not a source of silica for the chert. However, radiolarians are common components in the deeper allochthonous terranes of the spiculite belt.

Using conodont biostratigraphy, the spicule-rich chert units of the Phosphoria Formation can be directly correlated to those in the allochthonous terranes further north and west (Murchey and Jones, 1992). Much of the Permian chert of the spiculite belt in Canada is also associated with phosphorite deposits, so the oceanographic conditions required for phosphogenesis may extend far beyond the Phosphoria basin as well (Murchey and Jones, 1992).




Tosi Chert
Retort
Rex Chert
Meade Peak
L. Chert
Unit
50
40
30
20
10
5
M W P G Bc/Fl
Mud F M C Cong.
Tosi Chert
Retort
Rex
Meade Peak
Lower Chert
Unit
Samples
FC-21
FC-20
FC-19
FC-18
FC-17
FC-16
FC-15
FC-14
FC-13
FC-12
FC-11
FC-10
FC-9
FC-8
FC-6,7
FC-3,4,5
FC-2
FC-1
50
40
30
20
10
5
M W P G Bc/Fl
Mud F M C Cong.

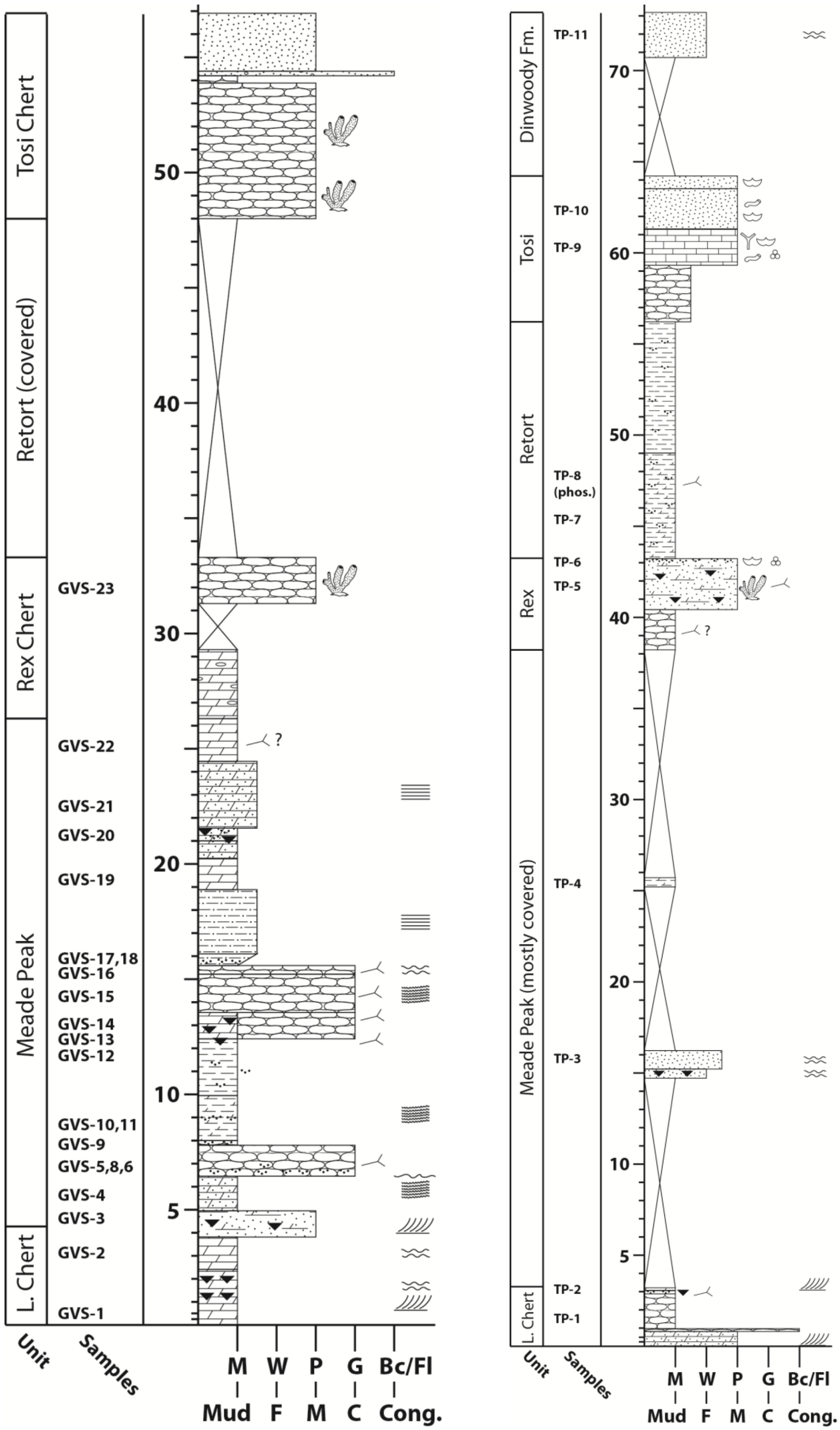

Tosi Chert
Retort (covered)
Rex Chert
Meade Peak
L. Chert
Unit
Samples
GVS-23
GVS-22
GVS-21
GVS-20
GVS-19
GVS-17,18
GVS-16
GVS-15
GVS-14
GVS-13
GVS-12
GVS-10,11
GVS-9
GVS-5,8,6
GVS-4
GVS-3
GVS-2
GVS-1
50
40
30
20
10
5
M W P G Bc/Fl
Mud F M C Cong.
Dinwoody Fm.
Tosi
Retort
Rex
Meade Peak (mostly covered)
L. Chert
TP-11
TP-10
TP-9
TP-8
(phos.)
TP-7
TP-6
TP-5
TP-4
TP-3
TP-2
TP-1
70
60
50
40
30
20
10
5
Unit
Samples
M W P G Bc/Fl
Mud F M C Cong.

Figure 4. Stratigraphic columns of logged sections in the Gros Ventre and southern Teton Ranges. A. Crystal Creek. B. Flat Creek. C. Gros Ventre Slide. D. Ski Lake. Key to symbols is included before A. See Appendix A for stratigraphic sections with unit descriptions, detailed locality information, and directions to localities.

GEOLOGIC SETTING

Lithofacies

In northwest Wyoming, the Phosphoria Formation and related deposits consist dominantly of chert, black shale, dolomudstone, and sandstone. Based on common occurrence of lithologies, fossil assemblages, and rare sedimentary structures, six main lithofacies were identified in the stratigraphy: *organic rich black mudstone and shale, peloidal phosphorite, spiculite, dolomudstone, dolomitic wackestone to packstone, and quartz sandstone.* Each facies is described below, followed by the interpreted depositional environment, discussed separately.

Organic-rich Black Mudstone and Shale

Organic-rich black shale is the dominant constituent of the two phosphatic shale members, and occurs in thick stratigraphic packages that contain dispersed phosphorite beds. In outcrop, beds are non-resistent, often fissile, and mostly form slope intervals. Outcropping beds are restricted mostly to just below more resistant overlying lithologies. The weathered surfaces of beds, where exposed, are mostly tan to light orange color, and blocky beds rarely exceed about 10 cm thick. Most of the black shale fizzes slightly with dilute HCl, indicating some component of carbonate. A few thick cherty mudstone beds are more resistant, crop out alone within a covered slope interval, and cannot be scratched with a rock hammer. The silicification of these beds may be due to increased density of sponge spicules in that part of the stratigraphy, as a source of silica.

There are no sedimentary structures or trace fossils recognized in the shale units at the localities for this study. Because of the fine grained fissile nature of most shales, no thin sections were made, so no microfossils were identified.

Depositional Environment:

Organic-rich shale is the dominant lithology within the Meade Peak and Retort Members in the deeper part of the basin, and its thick occurrence in western Wyoming has been interpreted as a marine transgression. The shale units have also been interpreted as accumulations of fine eolian silt, and therefore represent periods of limited carbonate production. The presence of minor amounts of carbonate within the shales in Wyoming however, indicates that even during highstands, carbonate production still occurred in the deep-water shale-dominated environment, though it was limited. The whole basin is characterized by input of fine eolian sediment, so the shale units represent periods of time when this was the dominant sediment input. Black shale is interpreted herein to be a deep water facies, down-slope from a carbonate ramp.

Peloidal Phosphorite (Fig. 5)

Phosphorite is defined as a sedimentary rock with a high enough content of phosphate minerals to be of economic interest. In northwest Wyoming, most phosphorites are coarse-grained and dominantly consist of phosphatic peloids, ooids, and bioclasts. Phosphorite occurs as thin, somewhat resistant beds typically within, or closely associated with black shale. Phosphorite is usually brown to dark gray color, and where weathered it is light gray to white. Phosphate grains are sometimes in a chert matrix, and often make up the base of a gradational bed, grading up into siltstone, sandstone, or conglomerate. Most phosphorite is present in thin beds within the black shale intervals, more commonly near the base.

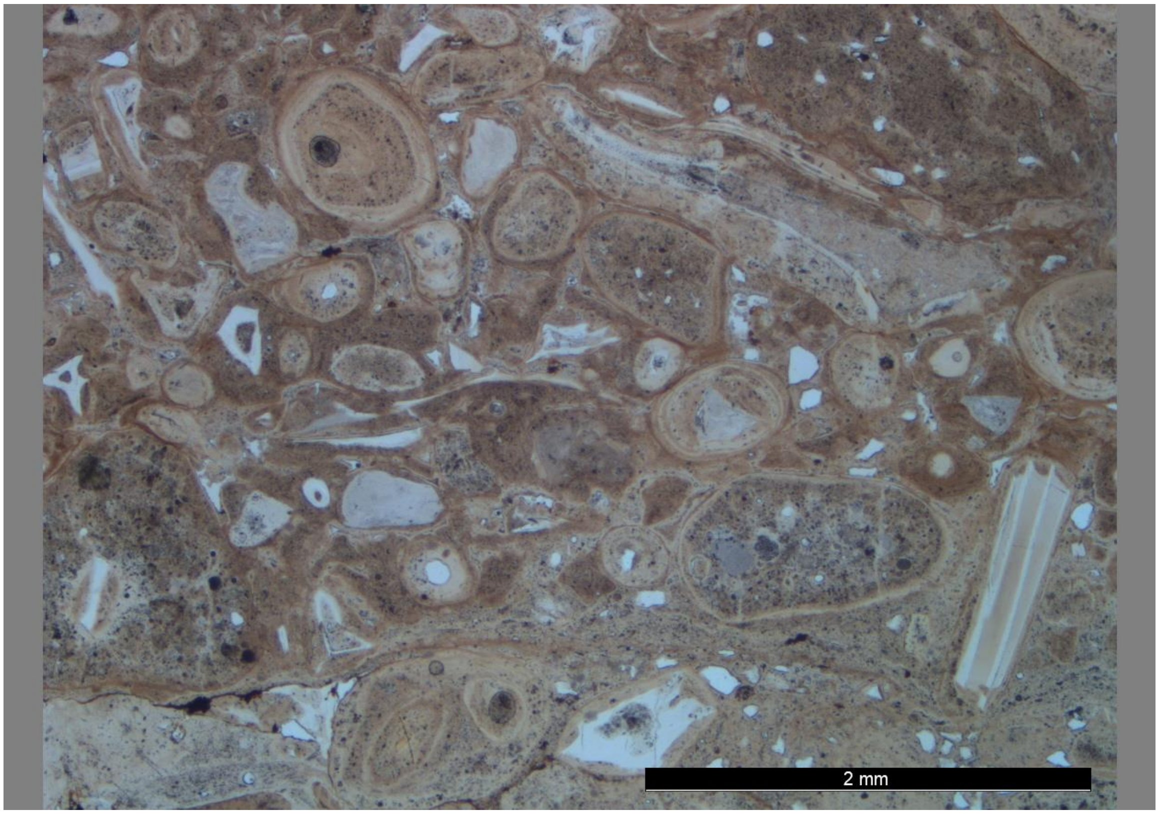


Figure 5. Photomicrograph of peloidal phosphorite in PPL from the Meade Peak Member at Gros Ventre Slide.

In thin section, the dominant sedimentary components are phosphatic ooids, peloids, and bioclasts. Bioclasts include fish scales and various gastropod and brachiopod fragments, preserved as carbonate fluor-apatite (francolite). Phosphatic ooids and other grains are often broken as well, and sometimes coated with subsequent phosphatic lamination. The cores of ooids are often small shell fragments, small aggregate grains, or what appears to be replaced micrite. In crossed polars, nearly the whole slide appears isotropic, because it is microcrystalline.

Depositional Environment:

Phosphorite deposits are present in nearly every part of the basin, and they are in a variety of forms, so phosphorite deposition was not likely restricted to one depositional environment. The dominantly ooid and bioclast-rich phosphorites present in the stratigraphy in northwest Wyoming seem to have formed in a high energy environment, relative to the rest of the stratigraphy. Ooids require wave action to form nicely concentric lamination, and broken ooids strengthen that argument.

The phosphatic fish scales that dominate some beds are in such anomalously high concentrations that they suggest periodic widespread death of fish in the Phosphoria sea. The basal bed for much of the Phosphoria exposures is the transgressive 'fish scale' layer, which can be correlated throughout the whole basin. These large-scale deaths are one piece of evidence that has been used to argue for marine upwelling, and anoxic conditions in the Phosphoria sea. Cold, nutrient-rich water from deep off-shore entered the basin, and allowed periodic algal blooms, which resulted in anoxic conditions for some time after. The fish scales represent one example of the organic matter that is deposited under anoxic conditions. Therefore, rather than being directly indicative of a certain depositional environment, the phosphorites likely record events. Their close association with the thick shale units however, has led previous works to interpret phosphorite as a deep-water facies. According to Pommer (2017), Phosphorites represent lowstands on the Wyoming shelf.

Spiculite (Fig. 6)

Spiculite is a name given here to any rock that is dominantly made of siliceous sponge spicules. Quantitative classifications were not used, but spiculite herein is at least half composed of spicules. It is often preserved as chert, but contains a wide variety of carbonatic and clastic sedimentary componenets as well. In outcrop, spiculites appear mostly as black bedded chert, in 10-40 cm thick beds, separated often by thinner wavy black shale interbeds. Within intervals of bedded chert, there are zones that contain distinct bedding-perpendicular cylindrical chert nodules made entirely of spiculite, which are described in their own section below. Spiculite occurs in nodule-like pockets as well, which are usually white or light gray in color.

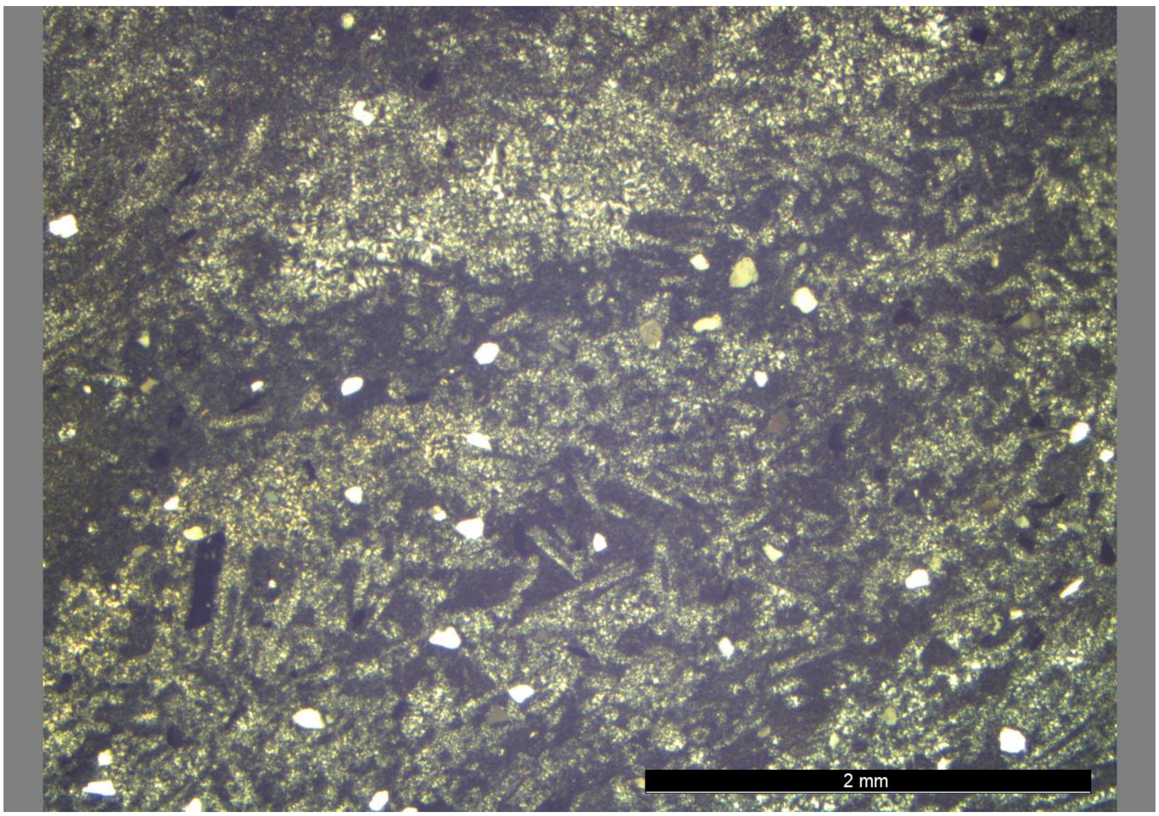


Figure 6. Photomicrograph of a chert spiculite in XPL from the Rex Chert Member at Gros Ventre Slide.

Depositional Environment:

Matysik et al. (2017) recognize two basic types of spiculites in Permian strata in Svalbard, light and dark-colored spiculites. Light spiculite contains less mud, and consists of coarser less well-sorted spicules. Most bedded chert is dark-colored, and probably represents muddy sediment, abundant in sponge spicules. Early diagenesis is responsible for silicification into bedded chert. The light colored spiculites are devoid of mud and may represent shallower water, where wave action removed the fine sediment. Overall, spiculite is probably largely deposited in a lower ramp depositional setting between carbonate-dominated sediment, and deeper-water black shale.

Chert is interpreted to be largely diagenetic, so the wide variety of spiculite appearances reflects a wide variety of spicule-rich sediment types, which were later silicified during diagenesis. Basically, siliceous sponges lived on a variety of substrates, and the high volumes of spicules that they produce are responsible for the chert.

Dolomudstone (Fig. 7)

Dolomudstone is usually thin to medium bedded, finely laminated, and is present in thick intervals within the measured sections. It is similar in appearance to the more resistant shale beds, and is tan to orange on weathered surfaces. On freshly broken surfaces, dolomudstone is usually gray, and fizzes in powdered form with dilute HCl. Dolostones are rarely bioclastic, but in some parts of the sections, they contain fossils up to densities of wackestones and packstones, which are given their own facies designation (see below). Large brachiopods and bryozoans, though rare, are the dominant skeletal components of dolomudstones. Chert nodules are common in dolomudstones, and appear as white to gray thin, wavy, discontinuous beds, or elongated wavy nodules. Some areas contain large (~5 cm) white calcite pockets, hollow in their center, lined with coarse calcite spar.

Most dolomudstones in thin section are fine to coarse recrystallized dolomite, so small fossils may be obscured. No fossils are recognizable in the thin sections analyzed for this study.

Fine angular quartz sand is dispersed in small amounts, and small brown phosphatic grains are common.

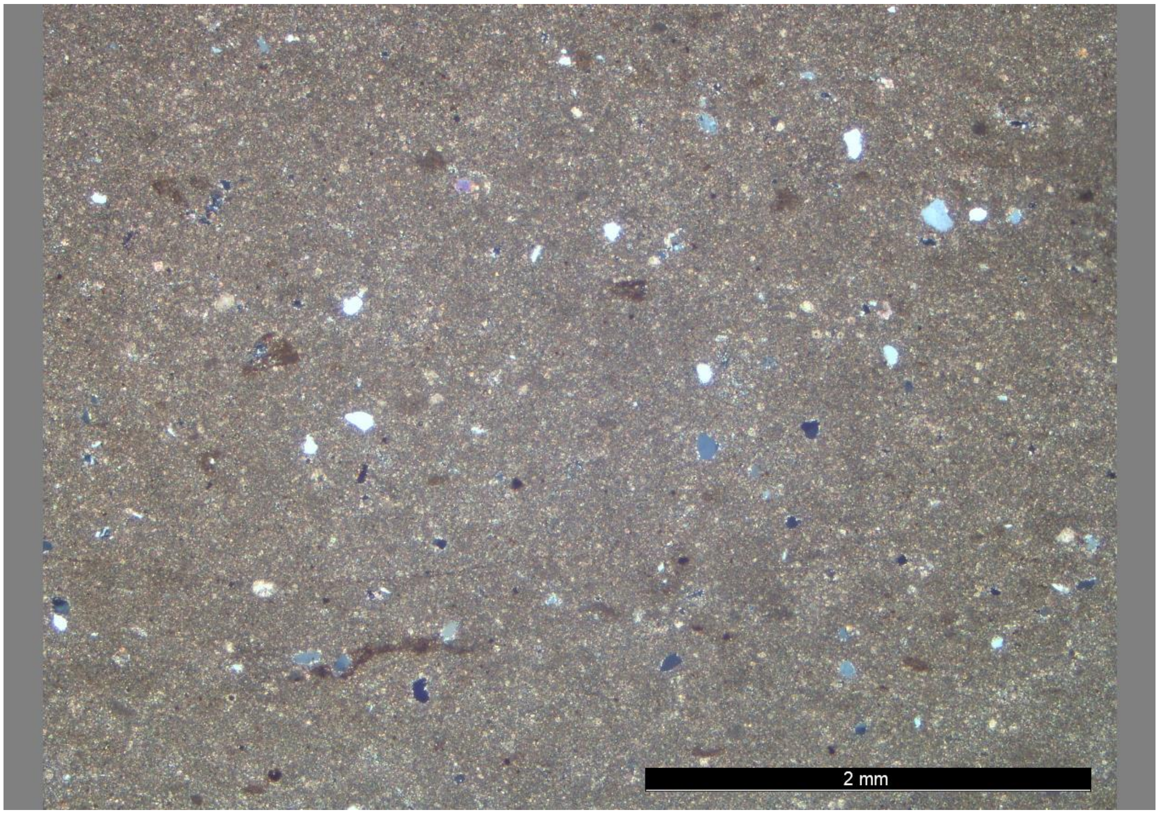


Figure 7. Photomicrograph of dolomudstone in XPL from the Meade Peak Member at Gros Ventre Slide.

Depositional Environment:

The lack of fossils and sedimentary structures are of little use for determining the depositional environment of dolomudstones in the exposures of Wyoming, but suggest deposition in deep water, below the effects of wave action. Given their stratigraphic relations with other lithologies, dolomudstone probably represents a distal carbonate ramp setting, down-slope from bioclastic limestones. Since there is usually only a minor fraction of siliciclastic sediment present in dolomudstones, they probably represent intervals of decreased eolian sediment input to the basin, allowing carbonate mud to accumulate.

Dolomitic Wackestone to Packstone (Fig. 8)

Bioclastic wackestone to packstone usually occurs in thick beds, but is not very common in the stratigraphy of northwest Wyoming. The common bioclasts are bivalves, large brachiopods, and large bryozoans. Gastropods and small cephalopods are less common. Cross beds are present in some examples, and vertical burrows are common including large *Diplocraterion*. This facies often contains fine sand, and is gradational with bioclastic sandstone facies. At Ski Lake, this facies is not preserved as dolomite, but as a calcite limestone.

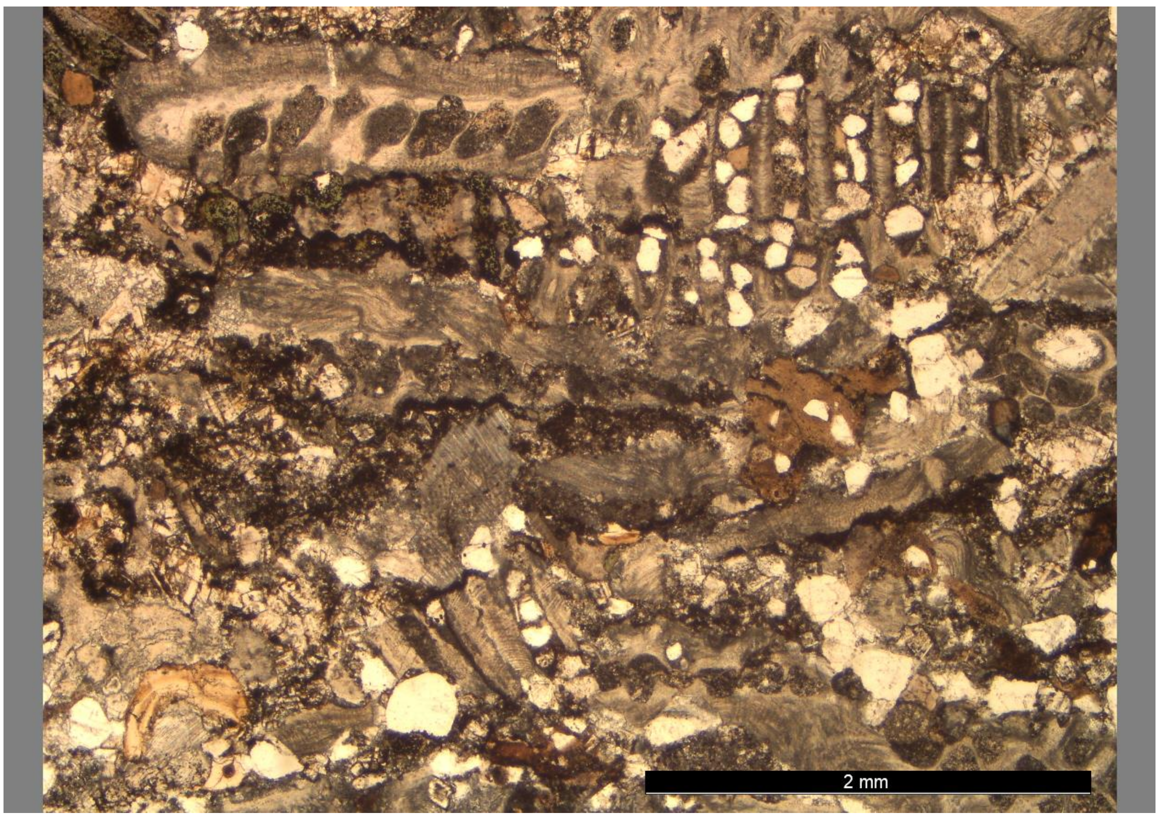


Figure 8. Photomicrograph of dolomitic packstone in PPL from the Retort Member at Ski Lake.

Depositional Environment:

Coarse bioclastic limestones are typically associated with a relatively shallow carbonate ramp environment. The presence of large carbonate-producing benthic organisms, burrows, and cross beds suggests a shallow water high-energy environment, such as a bioclastic shoal. The intact large bryozoans are indicative of a lower-energy, more protected environment, like a lagoon. Therefore, the dolomitic wackestone to packstone facies represents the shallowest carbonate ramp deposits in northwest Wyoming. It likely accumulated during periods of maximum transgression, when the carbonate ramp in central Wyoming prograded out (to the west) toward the deeper part of the basin.

Quartz Sandstone (Fig. 9)

Sandstone is common in the exposures in the Gros Ventre and southern Teton Ranges compared with other parts of the basin, and it forms thick, resistant beds. Sandstones vary from fine-grained quartz arenite to carbonate-cemented quartz arenite to sandy dolomudstone. Fine to medium angular quartz sand is the only clastic component of the sandstones. Microscopic sponge spicules are the most common fossils, though brachiopods and gastropods are common, along with large *Diplocraterion* burrows. Cross beds are rare, but recognizable in some beds, and no other sedimentary structures are present.

Depositional Environment:

It has been well accepted that the sandstones in the Phosphoria Formation are largely eolian deposits formed from pelagic accumulation of wind-carried sand (Carroll et al., 1998; Stephens and Carroll, 1999). Basinwide, sandstones are restricted to the northern and southern margins, and probably represent the larger fraction of eolian sediment carried into the basin. Fine silt and dust stayed in suspension and was able to be carried further and deposited in the central part of the basin, whereas the coarser fraction of sediment fell out nearer to shore. Although there are some cross beds, there is little evidence for channelized transport of sand in a subaerial or shallow near-shore environment. Instead, eolian proceses were probably mainly responsible for sand deposits in northwest Wyoming.

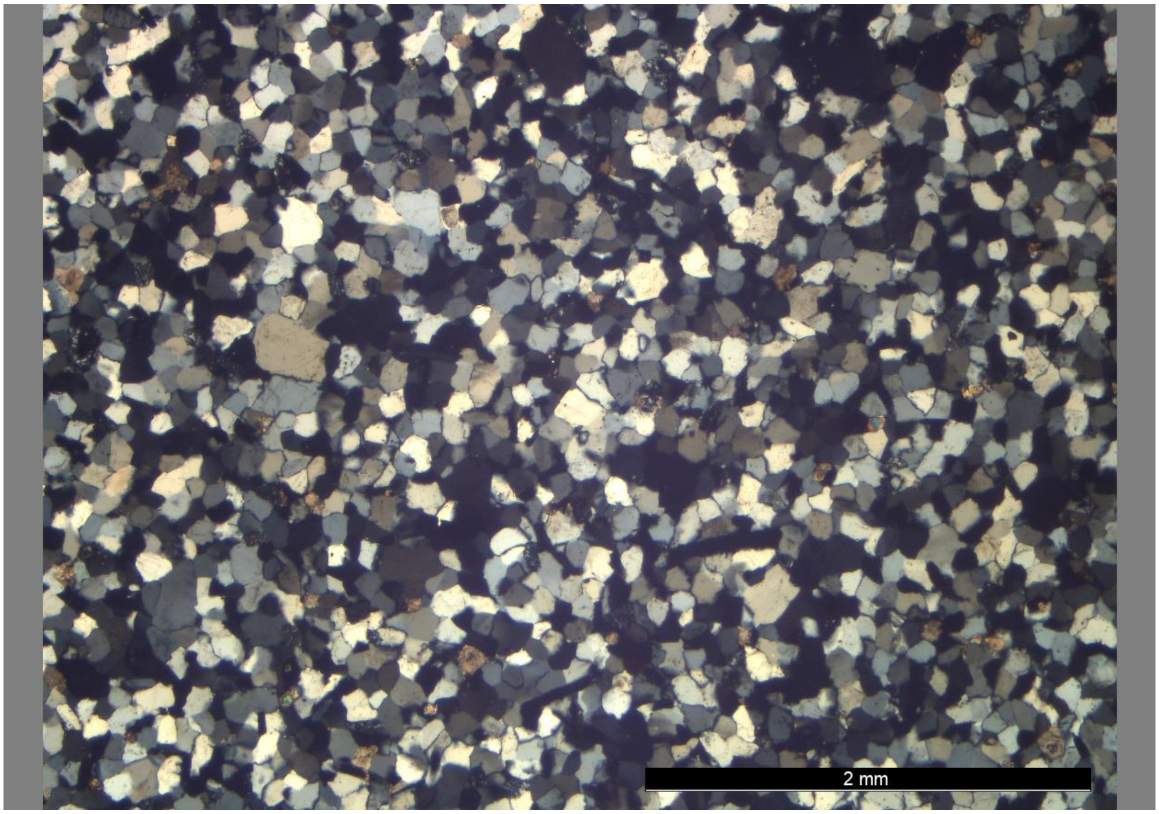


Figure 9. Photomicrograph of a fine-grained quartz arenite from the Lower Chert Member at Flat Creek.

CYLINDRICAL CHERT

Cylindrical chert nodules occur in abundance in both the Rex and Tosi Chert Members of the Phosphoria Formaiton in northwest Wyoming, and share many characteristics, though they have distinctly different appearances within the two units. In the Rex Chert (Fig. 15), they occur as irregularly cylindrical light gray chert nodules spaced out in a sandy dolomudstone to sandstone matrix. In the Tosi Chert (Figs. 10 and 11), they are more pronounced light gray chert cylinders that occur tightly packed within bedded black chert, where they make up entire beds.

The occurrence of these cylinders in the Rex and Tosi Members, and their morphology, will be described systematically below.

Dimensions

The cylindrical chert nodules exposed in the Tosi Chert Member are light gray, round to oval in cross section, and are about 5 cm in diameter on average, but range from 2 to 15 cm (Fig. 11A). The cylinders have very smooth walls, and retain the same diameter for their entire length. The axis is typically exposed for about 0.5 m length, but commonly can be traced as far as 1.5 m where well exposed (Fig. 10B and C). The (stratigraphic) tops of cylinders have never been observed, and the bottoms, where exposed (Fig. 11E), narrow slightly and round off at their base (Fig. 12A).

Cylinders exposed in the Rex Chert Member in northwest Wyoming are light gray, have irregularly circular cross-sections and are about 2-5 cm in diameter. They can be traced vertically for tens of centimeters, and have wavy, irregular walls (Fig. 15A, B, and E). On the top of bed surfaces, the cylinders are in positive relief, and are surrounded by halos weathered in negative relief for about 1 cm of the surrounding matrix rock (Fig. 15C). The cylinders appear to narrow slightly toward their bases (Fig. 15A), but this could be due to the outcrop surface obliquely transecting the cylinders. There are some rare examples where cylinders appear to branch upward as well, as in the base of the large cylinder in Figure 15B.

The cylinders in both the Rex and Tosi Members are made of recrystallized spiculitic chert, containing mostly large monaxon spicules, and some less common grains of angular quartz sand, silt, chert, and phosphorite (Fig. 16). The majority of the fill is chalcedony, obscuring some of the original cylinder fill texture.

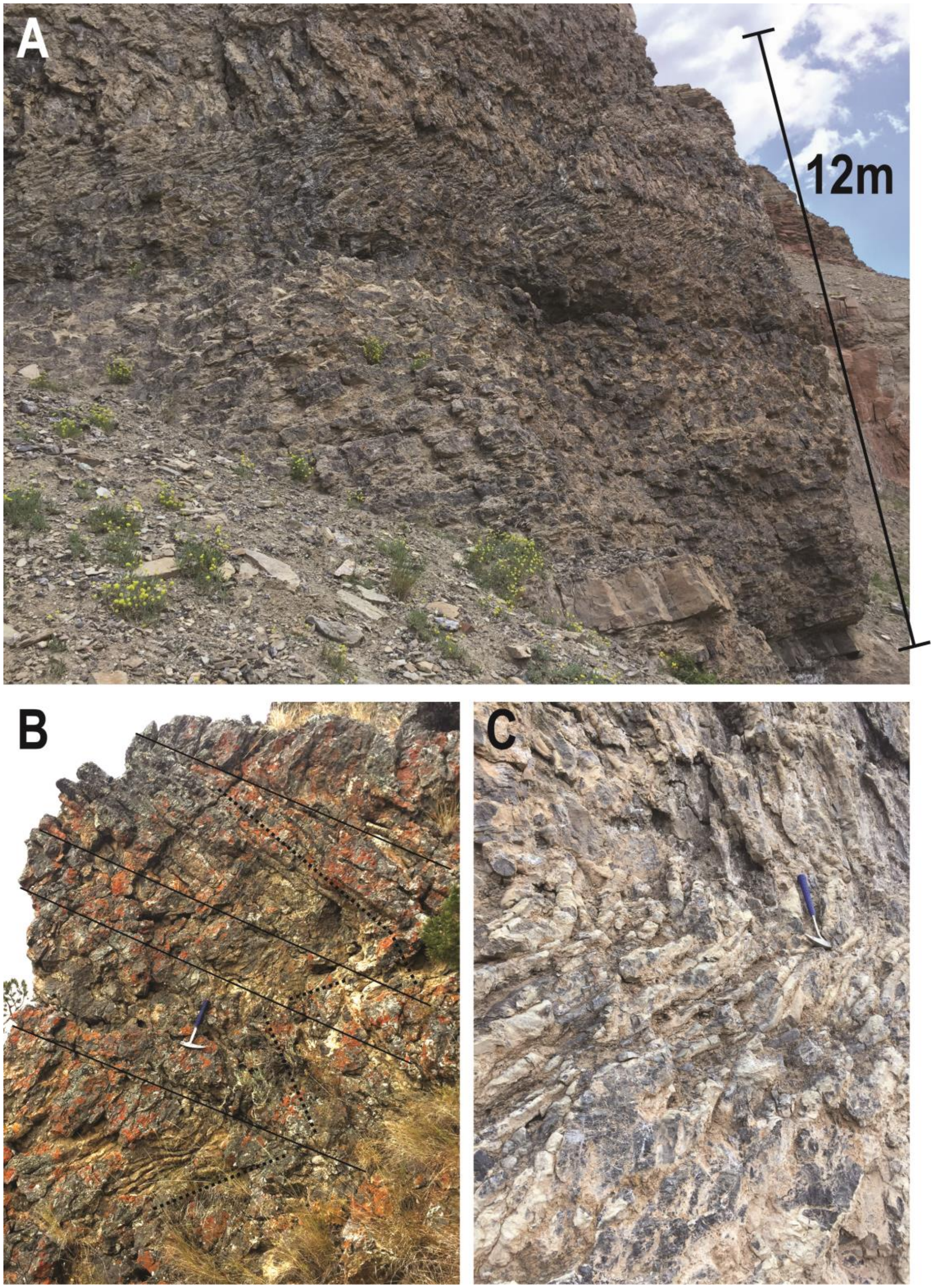


Figure 10. Outcrop photographs of the Tosi Chert at Crystal Creek (A, C) and Flat Creek (B). A. The full Tosi Chert Member at Crystal Creek. B. The upper half of the Tosi Chert at Flat Creek. Bedding is shown by solid black lines, and different zones of obliquely oriented cylinders are highlighted by dashed lines. C. A closer view of one of the oblique zones at Crystal Creek. Notice the transition into bedding-perpendicular cylinders above.

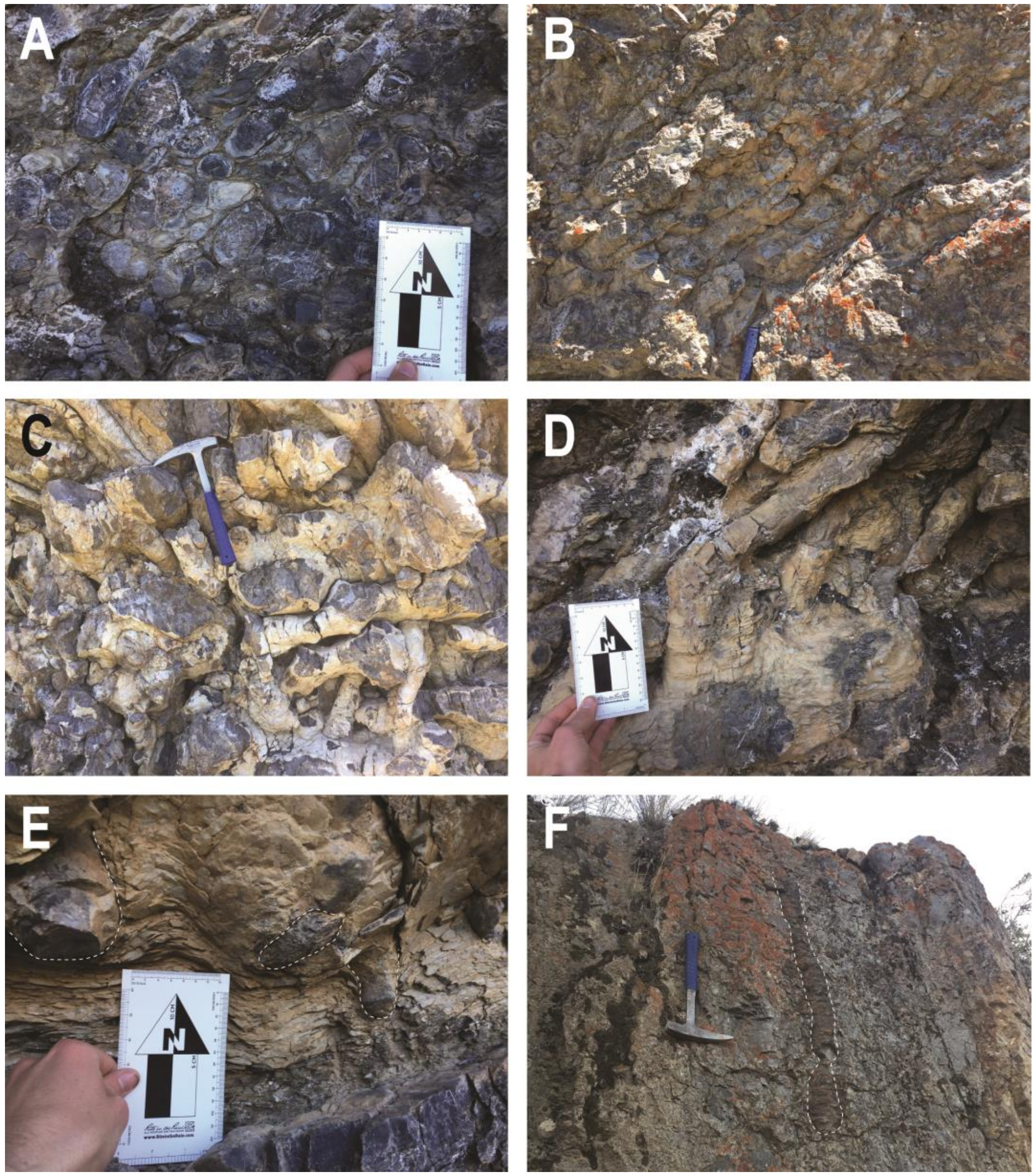


Figure 11. Outcrop photographs of cylindrical chert in the Tosi Chert at Crystal Creek (A, B, D, E) and Flat Creek (C, F). Rock hammer handle and scale card are parallel to stratigraphic up. A and B. Zone of tightly packed obliquely-oriented cylinders protruding out of the outcrop. C. Transitional zone between bedded chert of the lower Tosi and cylindrical chert. Irregular continuous chert beds contain cylinders oriented perpendicular to bedding, but cylinders can be seen bending between beds. D. Contact between a lower perpendicular zone, and an upper oblique zone of cylinders. One cylinder can be seen brittly bending. The lower, perpendicular part of the cylinder is wider and has cone-in-cone chalcedony rings in positive relief on its surface. The upper oblique part is smooth-walled and narrower. E. The base of a few cylinders (outlined) can be seen deflecting the muddy sediment downward, below a perpendicular zone of cylinders. F. A solitary sandstone cylinder, nearly one meter long, in a thick sandstone unit within the upper Tosi Chert at Flat Creek. The base narrows slightly to a rounded point, but the diameter remains constant for the length of the cylinder. The surface is irregularly wrinkled (rugose).

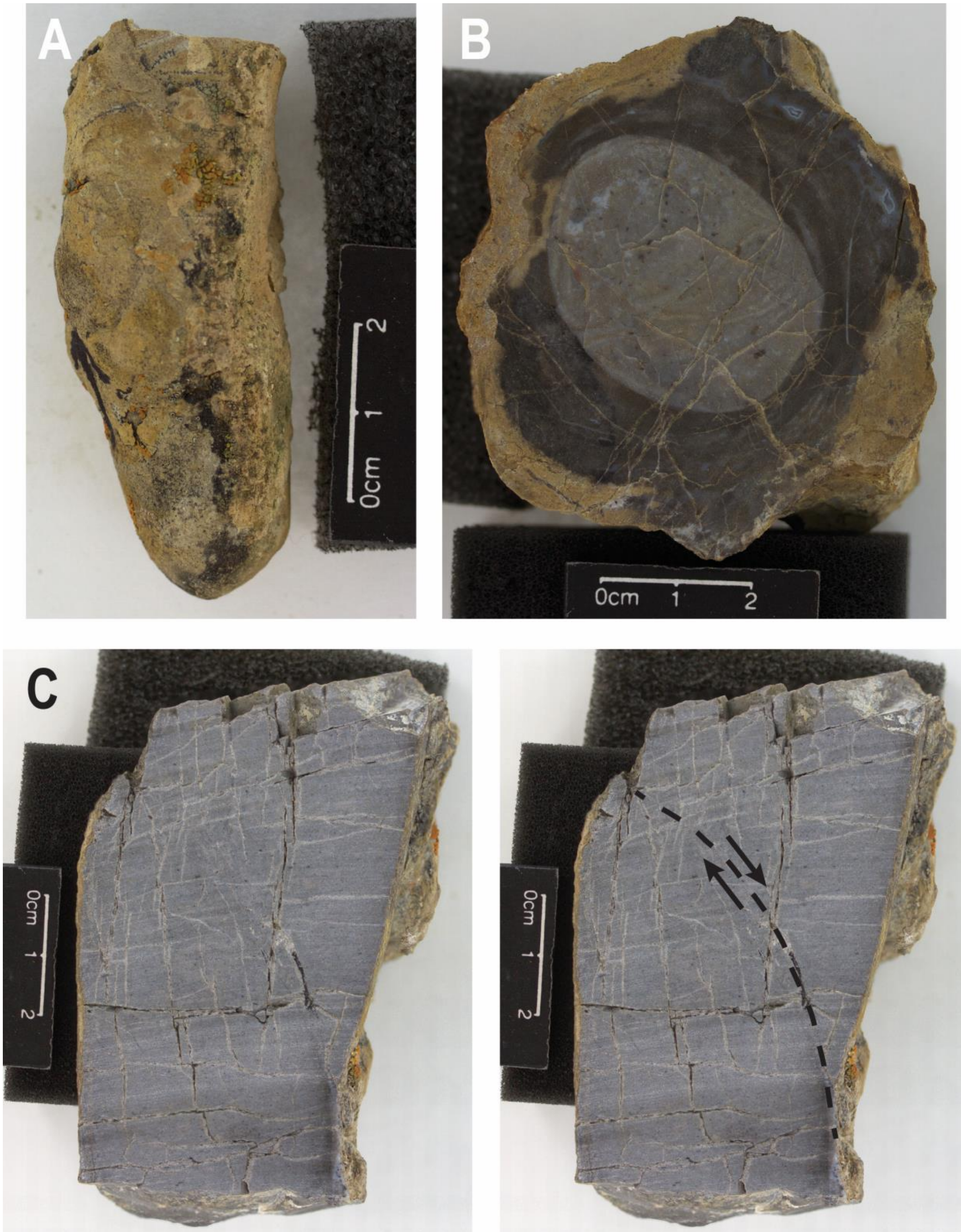


Figure 12. A. The base of a small cylinder, narrowing slightly and rounding off at the base. B. Lateral cut of a spiculite cylinder (central light colored oval), surrounded by silicified muddy sediment that contains no spiculites. C. Longitudinal cut of a cylinder (left), highlighting a small curved fault (right) with minor offset, indicating vertical shortening.

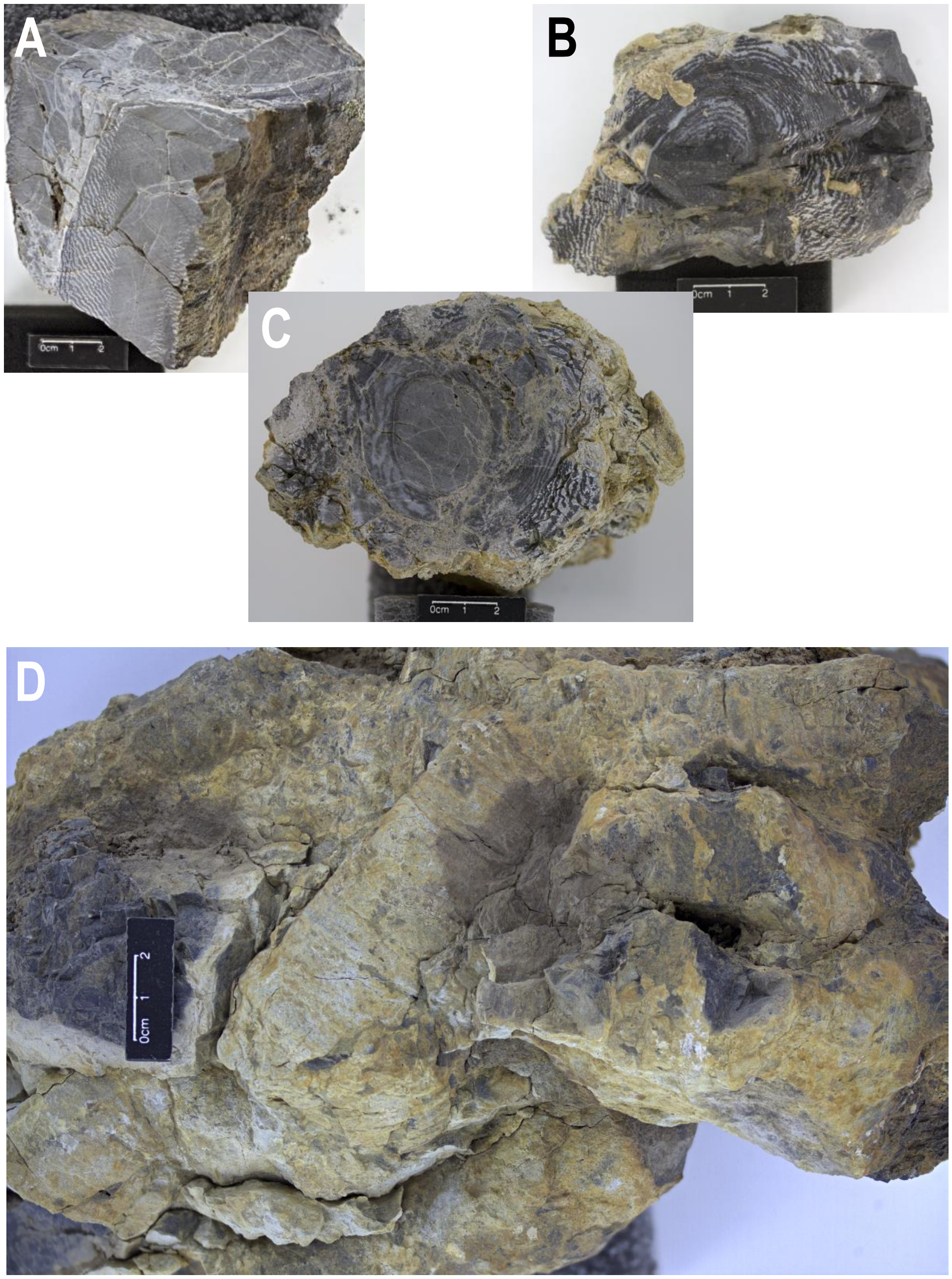


Figure 13. A. Longitudinal cut showing cone-in-cone chalcedony veins in cross section. Outside the cylinder is a cherty breccia. B. Oblique view of a cylinder showing the central column surrounded by cone-in-cone chalcedony veins in three dimensions. Notice how veins appear as parallel stripes on the front face, and can be traced in a curvilinear pattern around the central column on the top of the sample. C. Lateral cut of a cylinder surrounded by irregular tree ring-like chalcedony bands. D. Large sample containing at least four cylinders. One outlined cylinder has a rugose surface due to weathering in relief of cone-in-cone chalcedony bands. The cylinder is bent and highly brecciated due to vertical shortening or collapse. Chalcedony banding must predate bending of cylinder.

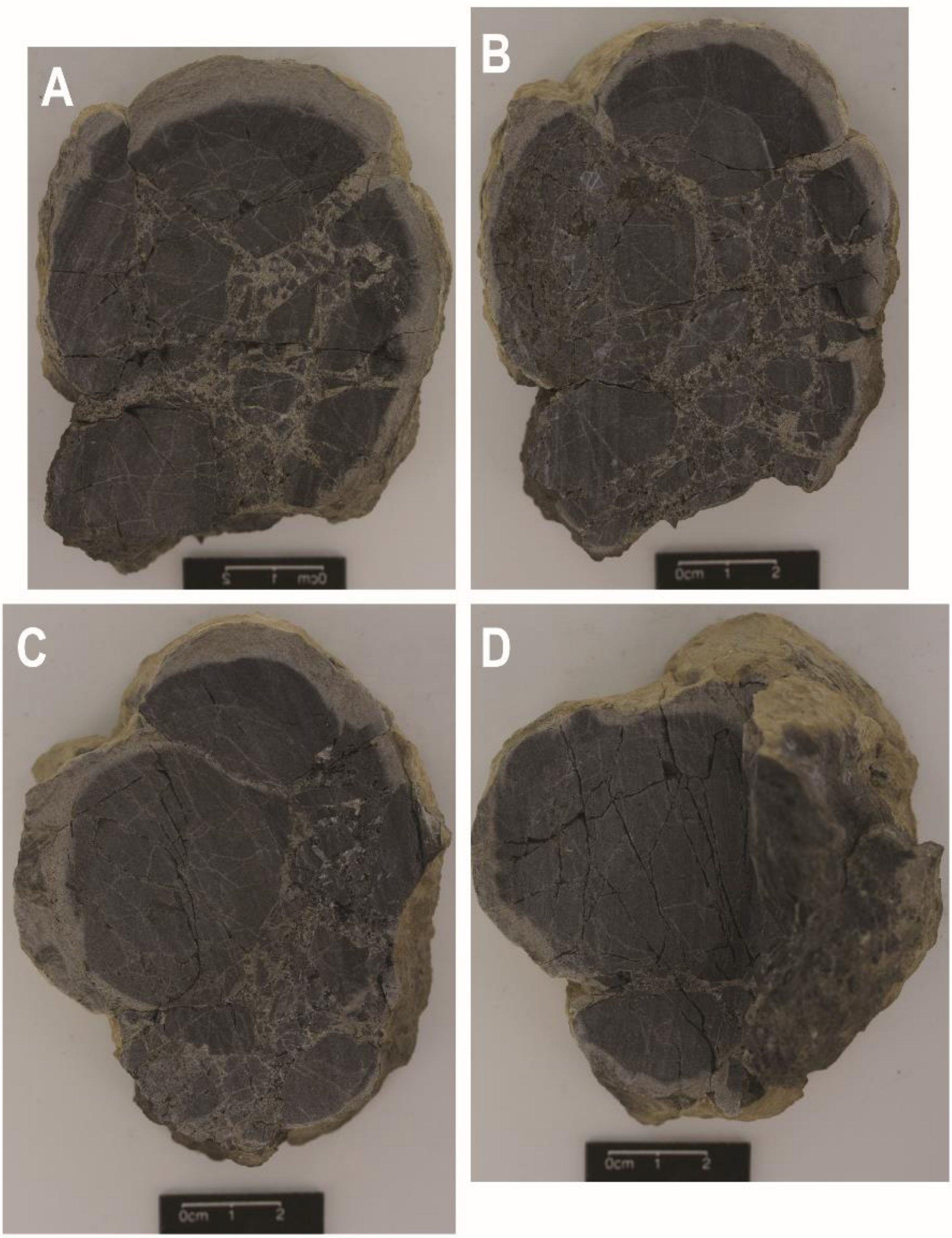


Figure 14. Four lateral cut slices through a single cylinder, which narrows toward its base. A is the uppermost slice, and B, C, and D are separated by 2 cm intervals. Notice the difference in brecciation moving from A-D, and that certain fractures can be observed crosscutting all slices. All scale bars 2 cm.

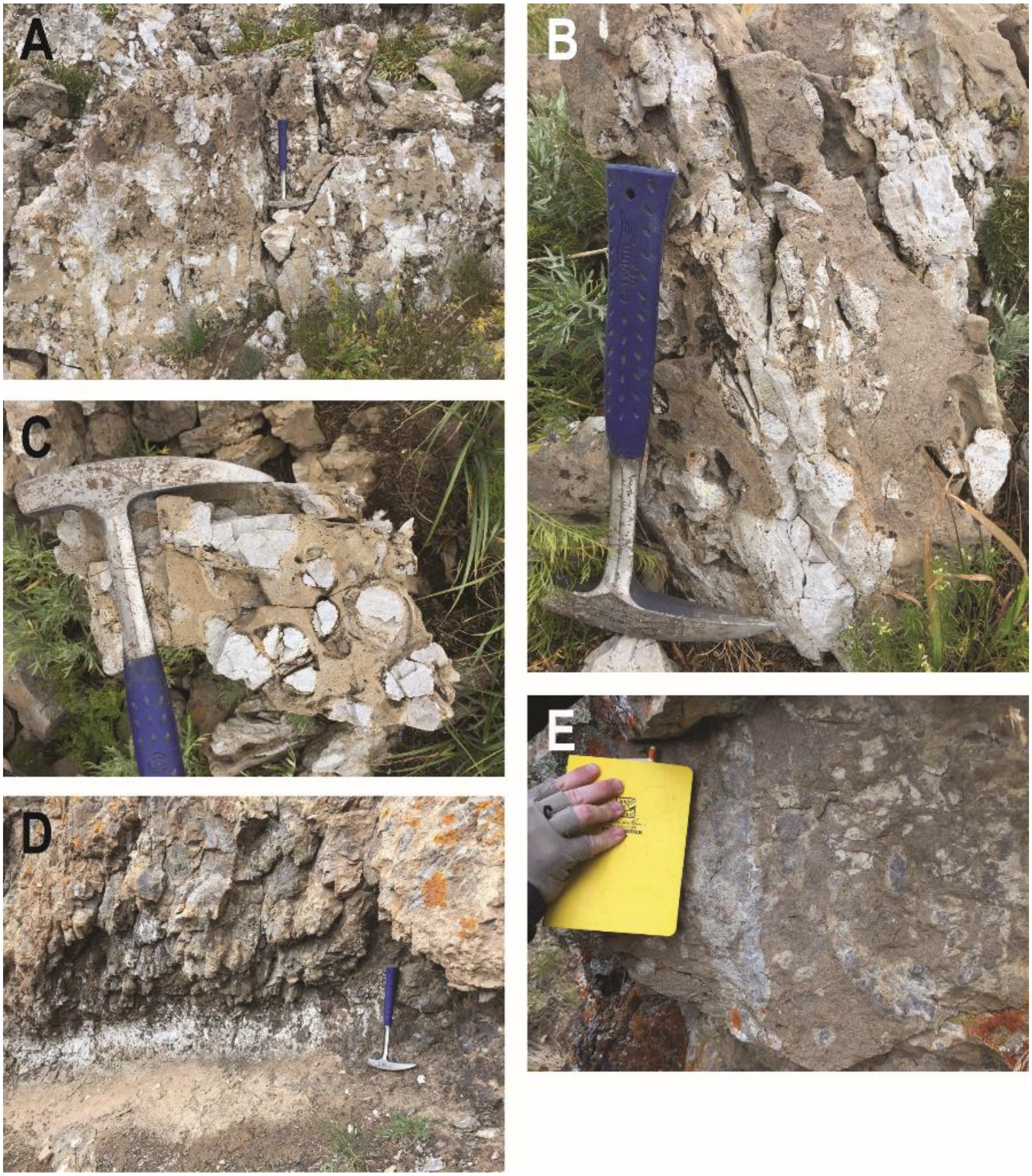


Figure 15. Outcrop photographs of cylindrical chert in the Rex Chert at Ski Lake (A, B, C), Crystal Creek (D), and Flat Creek (E). A and B. White chert cylinders in a cherty carbonate-cemented sandstone. Long axes are perpendicular to bedding, and chert is surrounded by a halo of carbonate spar. C. Top of a bedding surface showing roughly circular cross sections of cylinders, and carbonate-rich halos weathered in negative relief. D. Heavily weathered carbonate-cemented sandstone containing more-tightly packed, bedding-perpendicular cylinders, which weather in positive relief. E. Large gray chert cylinder and some smaller pockets of chert in a resistant sandy dolomudstone.

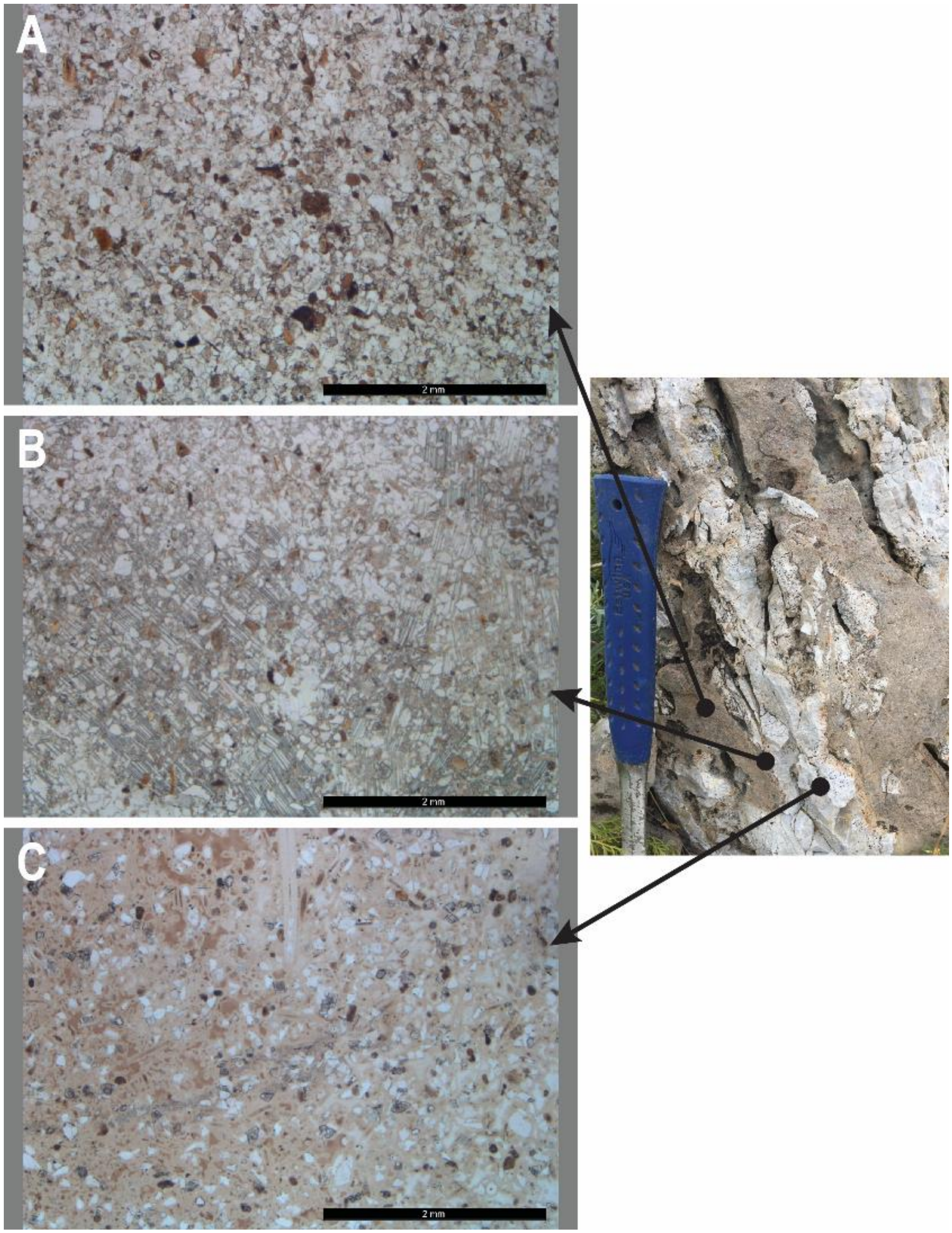


Figure 16. Photomicrographs in plain light of the Rex Chert at Ski Lake. A. Matrix between cylinders is a carbonate cemented quartz sandstone, with some sand-sized chert and phosphatic clasts. B. The negative relief halos encasing the cylinders are the same sandy sediment, but suspended in very coarse calcite spar. C. The interior of the cylinders are specular chert, composed mostly of sponge spicules, and with minor amounts of sand and phosphatic grains. All scale bars are 2 mm.

Orientation

Cylinders in the Tosi Chert are often oriented perpendicular to bedding, but in certain stratigraphic intervals, all of the cylinders are oriented at an oblique angle to bedding, tightly packed, and parallel to each other for their entire length (Fig. 10B and C). In the oblique zones, cylinders are oriented 30-45° to bedding, and orientation is consistent within each zone for very long distances. At Crystal Creek (Fig. 10A and C), fantastic exposure allows individual oblique zones to be traced for over a kilometer, and the orientation of cylinders remains remarkably consistent. Between different stratigraphic zones however, the orientations vary randomly in all directions (Fig. 10B).

The cylinders exposed in the Rex Chert are always observed perpendicular to bedding, and there are no oblique zones (Fig. 15).

Density

In many places within the Tosi Chert, the cylinders occur in such high densities that they are touching each other for their entire length, but contain no flattened surfaces at their contacts (Fig. 10B and C). They are so densely packed that there is rarely any matrix material filling the space between them (Fig. 11A and B). In bed-perpendicular zones, they are sometimes spaced out a few centimeters apart (Fig. 11E).

In the Rex Chert, the cylinders are spaced out a few centimeters to tens of centimeters apart from each other (Fig. 15A and C). On the top of bed surfaces, they appear to be randomly spaced out within the matrix rock, but are sometimes nearly touching (Fig. 15C).

Matrix

For the most part in the Tosi Chert, cylinders occur as tightly packed columns, with little to no matrix material between them (Fig. 11A and B). Where matrix material is preserved, it consists mostly of black chert, containing some fine quartz sand and mud, and minor amounts of carbonate material that crumbles when disturbed (Fig. 11E). The bases of cylinders can be seen in Figure 11E deflecting the fissile muddy dolomitic matrix sediment downward for up to 5 cm away from the cylinder. It is likely that outcrop exposure obscures the presence of matrix sediment, because the resistant cylinders are weathered in positive relief, and matrix material may have been removed from between them by surficial weathering.

The cylindrical chert-bearing beds in the Rex Chert are not as heavily silicified as the Tosi Chert, so the matrix is more discernible. At Ski Lake and Flat Creek, the matrix is calcite-cemented quartz sandstone, containing some sand-sized chert and phosphorite clasts (Fig. 16). The negatively weathered halo surrounding the cylinders is more carbonate rich, and consists mostly of quartz sand suspended in coarse calcite spar (Fig. 16). At Crystal Creek the cylinders occur in cherty dolomite, and the bottom of the bed is heavily weathered, so that cylinders can be seen protruding from the outcrop (Fig. 15D), similar to the Tosi Chert examples.

Stratigraphic Occurrence

The Tosi Chert is about 11 m thick where well exposed in the northern Gros Ventre Range (Fig. 10A), and is made dominantly of cylindrical chert. At Flat Creek for example, the lower 3.70 m is thin-bedded black chert with thin black carbonatic shale interbeds of the upper Retort Member, which disappear up-section, yielding only bedded chert (Fig. 11C). The overlying middle 6.0 m is made entirely of cylindrical chert nodules (Fig. 10B). Above is a recrystallized cherty sandstone for 1.9 m, which is overlain by another thin bed (0.35 m) full of smaller, but dimensionally proportional obliquely oriented chert cylinders within a sandy dolomitic matrix. The 6.0 m-thick middle sponge zone has few discernible bedding surfaces, but the orientation of cylinders varies with stratigraphic position (Fig. 10B and C), creating distinct laterally continuous stratigraphic zones. There are at least four zones within which all of the cylinders are oriented oblique to bedding (Figs. 11A and B). Above and below these zones the cylinders are oriented perpendicular to bedding, and can sometimes be seen bending (Figs. 10C and 11D), where they continue up into an oblique zone above. Overlying the uppermost cylinder zone at each locality is sandstone, or siltstone, often containing abundant brachiopods and bryozoans.

The cylinders in the Rex Chert are exposed only in one restricted interval, which at every locality overlies a thin interval of no exposure (likely shale). Below this thin recessive layer is usually a thick interval of dolomudstone and minor sandstone. Overlying the cylinder-bearing bed at every locality is fissile, carbonatic black shale and thin peloidal phosphorite interbeds of the Retort Member.

Alteration

All samples collected from outcrop consist of the light gray central round-oval cylindrical chert column, around which the morphological features vary. Along with cylinder orientation, the type of morphological features surrounding the cylinders generally correlates with stratigraphic positon. In the zones that are oblique to bedding, the cylinders are preserved mostly as just the central round to oval column, in some cases with a small amount of darker cherty material encasing them (Fig. 12B). Samples collected from these zones break off in segments typically no more than 20-30 cm long, but can be traced much further in outcrop. They often have slickenlines on their surfaces running parallel to the long axis of the cylinder, but show few signs of internal deformation. They are highly fractured, with chalcedony or dolomite-filled fracture sets oriented both roughly perpendicular and parallel to the long axis of the cylinders (Fig. 12C).

In zones within the Tosi Chert oriented perpendicular to bedding, the cylinders are more highly fractured, faulted, and are often wider overall (Fig. 11D), including the various textural features that surround the central column. Many of these samples contain various amounts of concentric chalcedony banding, which forms what have been previously described as cone-in-cone-like structures surrounding the central chert column (Figs. 13A, B and C). The chalcedony bands seen in a longitudinal cut are oriented between about 50° and 70° to the cylinder walls, have slight concave-up curvature, and the ‘cones’ open stratigraphically up (Fig. 13A). In lateral cut, the chalcedony bands appear as irregular concentric tree-ring-like stripes that are often truncated against the cylinder walls (Figs. 13B and C). On weathered surfaces of some cylinders, these chalcedony bands are weathered in positive relief, giving the cylinders a corrugated appearance (Fig. 13D). Some other samples contain large veins and pockets of microcrystalline quartz that seem to have expanded as they crystallized within fractures, and give the cylinders a brecciated internal appearance (Fig. 13A). Most of the samples contain many fractures, and some small faults that displace parts of the cylinders in vertical sense (Fig. 12C). Contained within some larger-diameter cylinders, there are small high-angle faults, nearly parallel to the long axis, which offset large slivers of the cylinders in a vertical sense, before chalcedony growth along the fault plane reattached them (Fig. 14). The resultant cylinders are wider, with a brecciated appearance in lateral cut, and concentric chalcedony bands that do not restore to an original shape (Fig. 14). Faults are not seen penetrating multiple cylinders in outcrop, but instead are confined within each. Some very large samples contain multiple cylinders that are bent and crumpled in spots, but remain mostly perpendicular to bedding (Fig. 13D).

Cylinders in the Rex Chert appear relatively undeformed, but in some places their walls are brecciated (Fig. 15B). They also contain fracture sets oriented both perpendicular and parallel to their long axis, but none show any fault offset.

DISCUSSION

Sponge Hypothesis

The cylindrical chert nodules described here have been interpreted in the past as burrows. Andersson and Sauvagnat (1998) assigned them the ichnofossil name *Skolithos grandis*, though no recent studies have focused on understanding the origin of these odd structures. Siliceous sponge body fossils have been suggested as a possible origin of these structures (Gutschick and Suttner, cited in Bromley et al., 1975), but studies lacked evidence to support the hypothesis. With new evidence presented in this study, the argument is made in the following sections that these cylinders are indeed siliceous sponge body fossils, and that they are not large burrows.

Abundant Spicules

Researchers have noted the abundance of siliceous sponge spicules in the chert-rich Phosphoria Formation for much of the last century (Sheldon, 1963; Rigby and Boyd, 2004), and it is well accepted that sponges were present in large volumes in the Phosphoria basin. In fact, siliceous sponge spicules are the principal allochems in the Rex and Tosi Chert Members, and their total volume in the Phosphoria Formation exceeds that of any other fossil group (Murchey, 2004). Spicules are common sedimentary components of the other members as well, and they are present in nearly all sedimentary facies. It is estimated that as much as two thirds, if not all of the silica within the chert members of the formation is sourced from siliceous sponge spicules (Yochelson, 1968; Bromley et al., 1975). The rarity of sponge body fossils in the Phosphoria Basin therefore, is in striking contrast to the abundant petrographic evidence for large populations of siliceous sponges on the sea floor of that time (Rigby and Boyd, 2004). Still, no sponge body fossils have been reported in the Phosphoria Formation, and only rare occurrences are noted in the Park City Formation.

The only fossils ever reported within the cylinders, which are always present, are siliceous sponge spicules. The spicules are randomly oriented and dispersed throughout each cylinder, and are in higher concentrations than the surrounding sediment (Fig. 16), where present. Another basic observation is that the sponges are preferentially silicified into cylindrical nodules, while the matrix rock is not. This is a result of the higher concentrations of spicules within the sponges, which act as preferred nucleation points for new silica growth. This is observed in the Permain spiculites in Svalbard, Norway too, where the 'silicification intensty' is controlled by the density of spicules within the sediment (Matysik et al., 2017). The formation of chert concretions occurs in the shallow subsurface during the transition from unstable opal-A to opal-CT to quartz, so presence of spicules (opal-A) in higher concentrations results in the formation of chert nodules after burial (Matysik et al., 2017). Therefore, even if the internal texture has largely been lost, it can be inferred that the cylinders contained a higher density of spicules at burial than the surrounding sediment. The best explanation is that the cylinders are sponge body fossils.

Sponge-like Morphology

The cylindrical central column preserved in all specimens resembles the columnar habit of the modern demosponge *Petrosia capsa* (Boury-Esnault and Rutzler, 1997). However, because of the poorly preserved internal texture of the cylinders, it is hard to determine if the sponges had an open central cavity, or spongocoel. Another possible modern analogue is the demosponge *Aplysina fistularis*, which is tube-shaped, with a large central spongocoel and open osculum at its top (Boury-Esnault and Rutzler, 1997). If the sponges had a spongocoel, it might be expected that spicules would be concentrated around the edges of the cylinder. However, spicules are randomly distributed within the cylinders, so either there was no spongocoel, or internal structures were lost after burial and silicification. The strikingly large size of the cylinders is also well within the observed size distributions for Permain fossil sponges in western North America (Rigby and Boyd, 2004; Rigby and Senowbari-Daryan, 1996).

The density of sponge bodies within the rocks is also reasonable, compared with modern sponge populations, which can occur in spectacular densities. These modern populations are composed of one or multiple species, and can cover up to hunderds of km^2 (Maldonado et al., 2016). The Southern Ocean around Antarctica is a modern example of an environment where sponges flourish and have dominated the benthic environments. Some species are as large as 2 m tall, and sponges occupy a broad variety of substrates, including sponge spicule mats 1 cm to 1 m thick (Maldonado et al., 2016).

Franson Member Sponges

The first discovery of sponges in the Park City Formation in Wyoming was by Finks et al. (1961), who found a small lithistid demosponge previously only known in the southwestern

United States. Rigby and Boyd (2004) reported small, intact twig-like siliceous sponges in a limestone bed within the Park City Formation in the Wind River Range, about 80 km southeast of the exposures in the Gros Ventre Range. These sponges are only a few centimeters tall at the most, but they are preserved in a similar style to the sponges described in this study, with a solid, fully silicified cylinder-shaped center. Some siliceous "irregularly cylindrical objects" at this locality were originally interpreted to be silicified burrows or worm tubes, but upon closer inspection by Rigby and Boyd (2004), were determined to be sponges that have undergone an "advanced stage of diagenetic destruction". The exact same argument is made herein, but for much larger cylindrical sponges, which have undergone an advanced stage of diagenetic destruction. Rigby and Boyd (2004) were able to conclude that these structures were sponges only after finding a few exceptionally preserved examples (the twig-like sponges) in the same rocks.

The Park City Formation represents the shallow-water carbonate ramp of the Phosphoria basin, and its Franson Member interfingers with the Phosphoria Formation in western Wyoming. The sponges recovered by Rigby and Boyd (2004) and by Finks et al. (1961) directly overlie the Meade Peak Member, and may be analogous to the Rex Chert sponges described herein, although they are much smaller, and occupied a shallower part of the basin. In the Rex and Tosi Chert, the much larger sponges likely underwent more intense silicification due to their close association with massive volumes of sponge spicules in the sediment, and they no longer preserve a recognizable internal morphology.

Mississippian Sponge Analogue

Large cylinder-shaped sponges have been documented in Mississippian limestones of the Arco Hills Formation at Leslie Butte in the Lost River Valley in south-central Idaho (Tapanila and Knecht, 2009). These sponges are preserved as chert within sandy bioclastic limestone and are very large, ranging from 10-35 cm in diameter and up to 2.5 m tall. Most are preserved perpendicular to bedding, while the rest are lying subhorizontal to bedding. The sponges consist of 3 to 5 concentric light-dark rings, with muddy sediment fill in the space between, corresponding to different intensities of silicification, surrounding a central spicule-rich zone. Some cylinders show evidence of branching upwards, but most are preserved as a cylinder shape only, narrowing toward their bases.

The large number of sponges observed by Tapanila and Knecht (2009) demonstrate the presence of a large-bodied siliceous sponge meadow in older Paleozoic limestones on the western margin of the North American craton. Their morphology and preservation may be analogous to the cylindrical sponges in the Phosphoria Formation. Those preserved perpendicular to bedding are interpreted to have been buried standing upright, while those lying on their sides were likely knocked over by storms. It is interesting that 2.5 m-tall sponges are preserved standing upright in limestone, because it is hard to imagine a mechanism for their burial that would preserve their delicate morphology, and not knock them all down. This is also an unresolved problem with the sponges in the Phosphoria Formation, but this Mississippian analogue demonstrates that burial and preservation in limestone is possible.

Poking Holes in the Burrow Hypothesis

Although these cylindrical chert structures have been given an ichnofossil name (Andersson and Sauvagnat, 1998), there are many aspects of them that are not consistent with an origin as burrows. First, the cylinders are very large, and their size and genesis as burrows cannot be explained (Bromley et al., 1975). No organism thought to be capable of making the burrows has been found preserved in them (Bromley et al., 1975), which is surprising considering how widespread is their occurrence. In contrast, the only fossils found in them, which are always present, are sponge spicules.

The size of the cylinders presents particular problems with their interpretation as burrows. Did an organism excavate over 1 m down below the sediment-water interface? According to Hein et al. (2004) the sedimentation rate was about 1 m/400-480 ky. Did the burrower live long enough for over 1 m of sediment to accumulate? Did multiple generations of the same burrowing organism occupy each burrow? The diameter of different cylinders is variable as well, which makes it unlikely that one burrowing species is responsible for all of them. Cylinders maintain a constant diameter for their full length, so it is also unlikely that an organism grew in size during burrow excavation.

The dense spacing of cylinders presents another dilemma. They have been observed and reported as composing up to 75% of the rock in some outcrops (Bromley et al., 1975), which would undoubtedly make burrows unstable and likely to collapse. Even if the burrower somehow stabilized the burrow walls, it is hard to imagine an ocean floor that contains more burrows than interstitial sediment, without burrows collapsing. Furthermore, no adjacent cylinder walls in the Tosi Chert are seen merging, and they are smooth, regular, and remain parallel to each other for long distances.

The infill of burrows is another issue. According to Sheldon (1957), there is no petrologic difference between the inside of cylinders and the surrounding sediment, except for the amount of microcrystalline silica present. Observations in this study suggest otherwise, and that the concentration of spicules is greater within the sponges than in the surrounding host sediment. It is likely in many cases that spicules were recrystallized to microcrystalline silica, and their grain boundaries are no longer apparent. Different sediment fill within the cylinders than the surrounding matrix cannot be explained by burrows, because there is no adjacent source for spicule-rich burrow fill sediment.

The host lithologies that contain cylindrical chert vary throughout the Phosphoria basin. The cylinders are seen preserved in sandstone, limestone, and chert, and often span multiple lithologies within the stratigraphic section (Bromley et al., 1975). In the Shedhorn Sandstone in Montana, single cylinders are seen spanning both sandstone and shale beds. Burrowing organisms prefer to live in certain substrates and certain energy environments, so it is unlikely that any one organism would burrow in such a variety of different sediment types.

Sheldon (1963) mentioned that the cylindrical (‘tubular’) chert occurrs in both the Rex and Tosi Members, and described them both in detail, largely in agreement with observations in this study. Sheldon (1963) also said that the internal structure of the cylinders is of no aid in determining their origin. Andersson and Sauvagnat (1998) did not mention the Rex Chert at all, and gave the name *Skolithos grandis* only to the cylinders in the Tosi Chert, and the Shedhorn Sandstone. This study is the first to describe both occurrences of the cylinders, and interpret them to be related in origin.

Life and Preservation

Previous studies documenting the cylinders in the Tosi Chert have assumed that they were originally perpendicular to bedding, regardless of their origin (Bromley et al., 1975; Scotford and Knecht, 1990; Andersson and Sauvagnat, 1998). Based on rare outcrop occurrences in the upper Tosi Chert of solitary sponges oriented perpendicular to bedding (Fig. 10F), this assumption is made in this study as well. In most outcrops however, the cylinders are deformed to oblique orientations and brecciated (Figs. 9 and 10). Their oddly-uniform oblique orientations have been interpreted as shear indicators (Scotford and Knecht, 1990), but the obliquity, along with small-scale faults, cone-in-cone chalcedony banding, and breccia within the cylinders, have not been explained satisfactorily. Observations from outcrop and hand samples, as well as observations documented in previous studies, are used here to develop a new model for the preservation of such structures.

The cylindrical chert nodules are not isolated to the Gros Ventre and Teton Ranges in Wyoming, but similar structures have been observed in Permian strata of Montana, Utah, and Nevada (Andersson and Sauvagnat, 1998). In chert and fine, dark, organic shale, the 'burrows' (cylinders) tend to be oriented oblique to bedding at low angles, and in sandstone, they tend to be perpendicular to bedding (Bromley et al., 1975). Furthermore, the external walls of 'burrows' are smooth in sandstone, and rugose, or wrinkled in mudstone and chert (Bromley et al., 1975). This is likely due to differential compaction and perhaps dissolution of host sediment. Quartz sand has little volume loss due to dewatering and compaction, so vertical shortening of these sponge-bearing units is minimal, and they retain a smooth-walled cylindrical shape. In shale, volume loss due to dewatering and sediment compaction is greater, and the cylinders have been vertically shortened. The whole thickness of the Tosi chert contains deformational features that suggest vertical shortening of the sponge-bearing strata, but the style of deformation differs between the oblique and perpendicular zones. This difference in deformational style is herein interpreted to be due to a difference in the matrix sediment in which the sponges are contained.

From basic field observations, it is apparent that sponges in the oblique zones had sediment between them upon deposition, and now much of that matrix material appears to be missing. The cylinders are touching one another for long distances, yet they don't have any flat edges to suggest that they grew together laterally to fill the space between them. If one restores the oblique cylinders back to a bedding-perpendicular orientation, they yield some empty space between them. If we assume, like all previous studies, that all of the cylinders were perpendicular to bedding, then there must have been some sort of sediment between them to allow them to crystallize as solid chert cylinders. Since the host sediment is not apparent, it is likely to have either been carbonate, which dissolved and was mobilized away at some stage of diagenesis, or muddy sediment that has been weathered away from outcrop surfaces, leaving only cylinders exposed in positive relief. Compaction or loss of matrix material would result in tilting and collapse of cylinders to fill void space, and a large amount of vertical shortening after burial. Restoring the oblique cylinders back to a vertical orientation yields a shortening magnitude of about 30-50%.

Cylinders that are observed in bedding-perpendicular zones are usually wider and have irregular walls compared with those in oblique zones. Further, they contain cone-in-cone chalcedony bands and small high-angle faults. These structures also indicate vertical shortening of the cylinders after deposition. Once again, it is assumed that these cylinders were perpendicular to bedding upon deposition, and that they were the width of their light gray central cylinder. Cone-in-cone chalcedony bands fringing the central column are interpreted to accommodate widening of the cylinders during vertical shortening, under a greatest principal stress parallel to the long axis of the cylinders. The two lesser principal stresses, if equal to each other, would confine each cylinder in all lateral directions, so that no preferred directional weakness is present, and chalcedony-filled veins opened up in a circular pattern within and around the central column. This stress field could be generated by vertical compaction of cylinders in confining sandy or carbonatic sediment. There are also some examples of high-angle faults (Fig. 12C), and complete offset of parts of cylinders adjacent to themselves (Fig. 13D, 14). These structures are discontinuous between adjacent cylinders, indicating that the host sediment was not lithified at the time of fault movement, or that there has been volume loss of the host sediment. The faults indicate vertical shortening of the cylinders, and overall widening by emplacement of faulted pieces into the interstitial space between each cylinder. As a result, sponges in the bedding-perpendicular zones are often wider and more fractured than in oblique zones.

Matysik et al. (2017) used oxygen and carbon isotopes, along with detailed textural analysis of spiculites in Svalbard, in order to constrain their diagenetic history. Because of striking similarities in the stratigraphic succession between those deposits and the Phosphoria Formation, the various stages of mineralization described by Matysik et al. (2017) are applied to the stratigraphy in this study. In a simplified summary, the transition from opal-A to opal-CT and chert nodule growth takes place in the shallow subsurface. Mechanical compaction of the sediment occurs until medium burial depths, and fracturing, brecciation, and chemical compaction take place at medium to deep burial depths. The majority of silica crystallization occurs at shallow to deep burial depths, and the minor calcite and dolomite crystallization occurs at deep burial depths.

A New Model

Based on deformational features observed throughout the Tosi Chert, and a model for chert burial and diagenesis in the time-equivalent Tempelfjorden group in Svalbard (Matysik et al., 2017), I have generated a new model for the preservation of these sponges, illustrated in Figure 17.

Sponges were growing in close proximity to one another on the sea floor, but spaced out slightly more than they are presently observed in outcrop. One unresolved issue is that there is no suitable mechanism to bury 1 m tall sponges without knocking them over and disaggregating them. There are three possibilities that seem most likely: the sponges were buried by rapid event sedimentation and any remaining unburied portions were truncated above the sediment and disaggregated into spicules; the sponges were being actively buried as they grew upwards, and may not have been taller than a few centimeters off the sea floor at any given time; or the sponges were rigid bodies, or were silicified to rigid structures before burial, and sediment accumulation slowly buried them. The data in this study is not sufficient to answer this question, but I tend to favor the latter, that there was continued sponge growth during burial, because there is no evidence of a break in slope on which sediment gravity flows could occur, or of rapid event sedimentation elsewhere in the section.

Cylinders have been reported in the Shedhorn Sandstone up to 6 m long (Bromley et al., 1975), which also leads to the interpretation that sponges were growing as they were being buried. For the purposes of this new model (Fig. 17), the growth and burial mechanism could be any of these and the result is the same. To not rule out any options, the model in figure 17 includes both sponges that are buried and then truncated at the sediment surface, as well as sponges that continue to grow during sedimentation.

Following initial burial, the sponges were preferentially silicified during the opal-A to opal-CT to quartz conversion in the shallow subsurface and became solid cylinders of chert before the surrounding sandy sediment was lithified (Fig. 17). During and after the sponges were silicified, and with continued burial, vertical compaction of the sediment resulted in buckling of sponges contained in certain beds. In more confining beds, faulting, cone-in-cone vein mineralization, and brecciation of cylinders took place (Fig. 17). After shortening, the remaining matrix rock was fully silicified to chert. Finally, after even deeper burial, dolomite precipitated from pore waters along grain boundaries, and filled small fractures, as the latest stage of diagenesis.

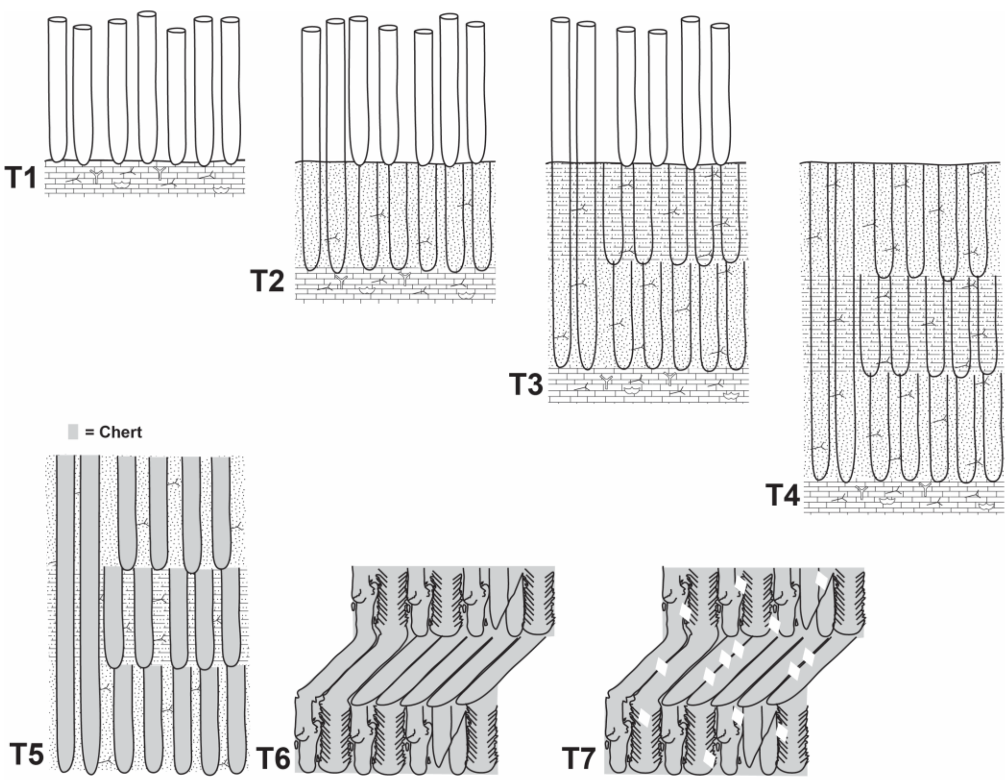


Figure 17. Model for preservation and vertical shortening of sponges. T1-T4 show three successive generations of sponges occupying the substrate and then being buried. Notice the leftmost two sponges are shown continuing to grow during burial. T5: after burial to some relatively shallow depth, sponges get preferentially silicified into solid chert cylinders, while sediment remains unlithified. T6: vertical compaction of sediment causes brittle deformational features to form, with separate zones of tilting, and zones of faulting, cone-in-cone bands, and brecciation. T7: late stage dolomite growth along grain boundaries and fractures.

Are They Useful Shear Indicators?

Scotford and Knecht (1990) interpreted the zones of obliquely-oriented cylinders within the Tosi Chert as pressure solution strain markers in the Gros Ventre Mountains. They argue that two stratigraphic zones of ‘bent tubular chert’ contain cylinders that are consistently plunging to the northeast across a broad region, and record flexural slip during the folding of the Gros Ventre anticline. Observations by the present author at many of the same outcrops, and the photograph in Figure 10B however, reveal that there are more than just two zones of obliquely oriented cylinders, and that their orientation varies greatly between zones. Although trend and plunge measurements were not made in this study, one can clearly see flaws in the interpretation by Scotford and Knecht when looking at outcrop exposures. Scotford and Knecht (1990) also use the truncation of cylinder walls as evidence of pressure solution within the cylinders during deformation. This truncation is interpreted herein as small-scale faulting (discussed in a previous section) offsetting parts of the cylinders in a vertical sense, and giving them a truncated appearance (Fig. 14).

The model proposed by Scotford and Knecht (1990) fails to address why the cylinders are so tightly packed in the oblique zones, and why the shear strain was confined only to two discrete stratigraphic zones when cylinders are not. Also, there are many studies (eg. Bromley et al., 1975; Andersson and Sauvagnat, 1998) that document the same bending of these cylinders in outcrop exposures that reach far outside of the Gros Ventre anticline, into Montana (Fig. 18).

In this study, the tilting is interpreted to be a result of dewatering and compaction, and possibly partial dissolution of matrix sediment between the solid chert cylinders. Due to a volume loss of matrix, the unsupported cylinders in those zones tilted by sliding against one another, and collapsed. Within each oblique zone, the cylinders all tilted the same direction during vertical compaction, but each zone seems to have tilted in a random direction.

Regional Occurrence

The cylindrical chert within the Tosi and Rex Chert members has been observed by the present author at three localities in the Gros Ventre Range (Crystal Creek, Flat Creek, and Gros Ventre Slide), and one locality in the southern Teton Range (Ski Lake) in northwest Wyoming. These structures have been reported from previous studies in both the Rex and Tosi Chert Members throughout northwest Wyoming (Sheldon, 1963; Scotford and Knecht, 1990). Furthermore, similar structures with varying forms of preservation have been reported in Permian chert and sandstone units throughout most of the shallow-water extent of the Phosphoria Basin in southwest Montana, western Wyoming, northern Utah, and northeast Nevada (Fig. 18) (Bromley et al., 1975; Andersson and Sauvagnat, 1998). In Montana, the cylinders occur within the Tosi Member and the Shedhorn Sandstone, and in northwest Utah and northeast Nevada, the cylinders occur in the correlative Gerster and Murdock Mountain Formations. They are reported as being up to 6 m long at some localities, and occur as chert cylinders in a chert matrix, sandstone cylinders in a sandstone matrix, or as sandstone cylinders in a chert matrix (Andersson and Sauvagnat, 1998). In northwest Wyoming, cylinders occur only as chert, and are preserved in a chert, sandstone, or sandy dolomite matrix.

Although the host lithologies, preserved forms, and density of the cylinders varies between localities, first order observations indicate that they all appear to be genetically related structures (Bromley et al., 1975; Andersson and Sauvagnat, 1998). Also, beds that contain the cylinders are observed in a definite part of the section at all localities (Sheldon, 1963), suggesting that they are syngenetic in origin.

Location of the sponge communities within the basin likely depends on conditions other than substrate, because they occur in a variety of facies. The occurrence of sponges within the basin follows a semicircular pattern shown in Figure 18, surrounding the central, deepest part of the basin. Sponge spiculites also occur in a semicircle, basinward from the carbonate ramp, and therefore probably represent an extensive array of sponges living in that part of the basin. There is no doubt that sponges lived throughout most of the basin for certain time intervals (Sheldon, 1957; Rigby and Boyd, 2004), but these exceptionally preserved large-bodied sponges probably represent a unique depositional environment where they thrived within the basin.

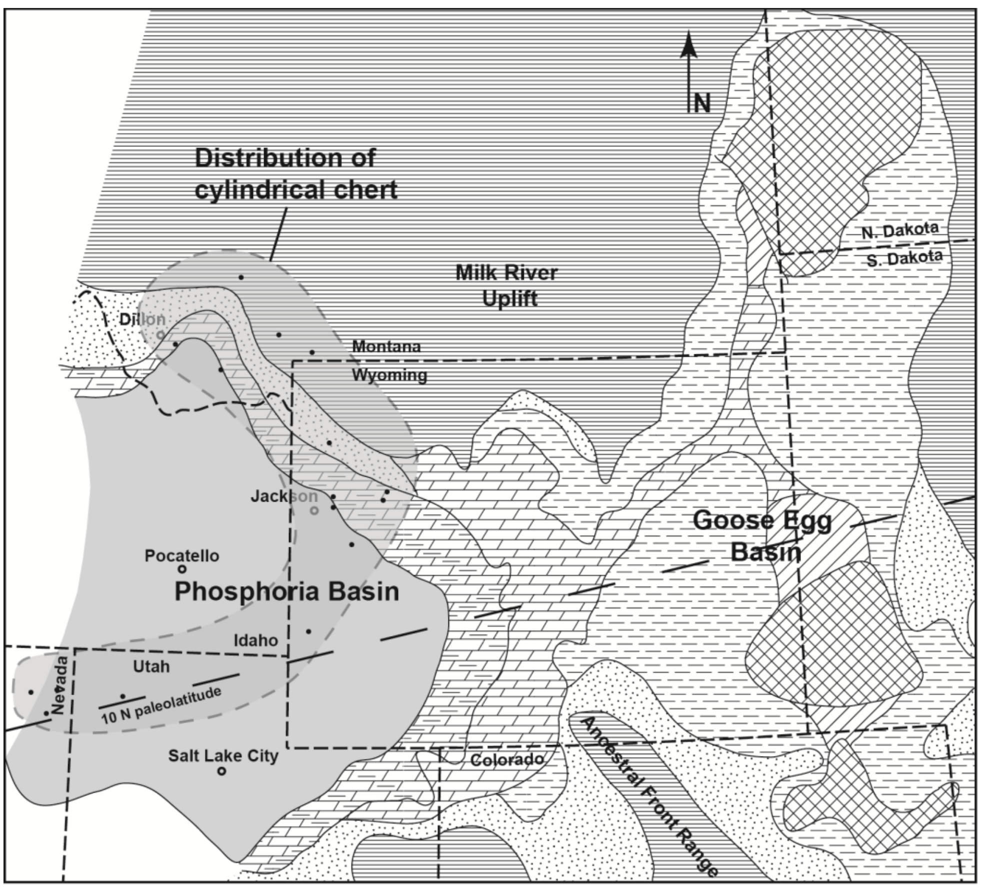


Figure 18. Map of observed cylindrical chert in outcrop within the Phosphoria Basin. Points are from Andersson ad Sauvagnat (1998).

What do Sponges Indicate about Paleoceanography?

The most common spicule types in the Phosphoria Basin are large monaxons which do not branch, and less common are rhaxes, strongyles, and spheroidal microscleres. Together, these make up a distinct fauna of sponges in Permian western North America (Murchey, 2004). In the thin sections from this study, monaxons are the overwhelmingly dominant spicule type, and branching spicules are not recognized. These abundant monaxons are thought to be demosponge spicules, and hexactinellid spicules, though often present, are a minor fraction (Murchey, 2004). Based on a regional analysis of spiculites within the Phosphoria Basin, the sponges presented here are interpreted to be demosponges, rather than members of another class.

Samples from the Rex Chert in Idaho (see Murchey, 2004) often contain broken, current-oriented spicules, sometimes with enlarged axial canals suggesting transport and partial dissolution before deposition. Furthermore, the spicules are interpreted to be from sponges living at less than 50 m water depth (oral communication, cited in Cressman and Swanson, 1964). In a model by Murchey (1990), the ratio of sponge spicules to radiolarians correlates inversely with increasing depth of living position. Based on this model, it is inferred that sponges in the Phosphoria formation were living at depths of less than 150 m (Murchey, 1990). Interbeds of coarse bioclastic limestone strengthen this argument for a relatively shallow, perhaps lower ramp environment.

No radiolarians are present in the Phosphoria Formation, but they occur in the Havallah basin of Nevada. The presence of radiolarians and general lack of sponge spicules in Permian rocks further outboard of the Phosphoria Formation (Murchey, 2004) suggests most western chert formed in a deeper water environment, where sponges are rare. On a regional scale, spicule-dominated deposits only occur in the Phosphoria Basin, and in isolated shallow subbasins to the southwest, coincident with Phosphorite deposition.

Sponge abundance is often strongly linked to food supply. Sponges are filter feeders, and lack central digestive organs, so they feed mostly on the finer fraction of particulate matter available in the water column (Finks, 2009). Their food includes molecules on the smallest level, organic detritus, bacteria, small protozoa, and single-celled algae. Particles of this size occur in all oceanic waters, and therefore so do sponges (Finks, 2009). Furthermore, the highest density sponge populations in modern oceans are in areas with high organic productivity, and food availability. In most cases, they outcompete other benthic organisms for substrate space, and are the only organisms living in certain environments. Examples include concentrations of sponges near the mouth of rivers in Bermuda carrying high amounts of organic detritus, or off the North Cape of Norway, where converging ocean currents yield large kill-offs of plankton, and sponges constitute 90% of the local biomass (Finks, 2009).

In the Phosphoria sea, the sponges likely outcompeted brachiopods and bryozoans for both space and for food in the water column, and they seem to have occupied the entire seafloor in certain areas for periods of time. The sponges also may have thrived in conditions not suitable for their competitors, giving them an advantage. Fossils of other benthic organisms are lacking where sponge spicules are most common. The high amounts of organic matter raining down through the water column in the Phosphoria sea would have been an excellent food source as well, to support the massive populations of sponges observed in the Tosi and Rex Chert Members.

According to Finks (2009), the morphology of sponges depends largely on availability of food, and the amount of energy in the water column. Branching sponges are often strong, and have a high amount of surface area to filter food from the water column. Thin-walled tube-shaped sponges are mechanically weaker, and occupy low-energy environments, but still maintain a high surface area relative to size of the sponge body.

In the Phosphoria Sea, the sponges described herein are likely columnar or tube-shaped, and do not branch, suggesting that they lived in a low energy environment. The abundance of food in the water column allowed them to thrive in large populations. Sponges feed on suspended food filtered from water far above and beside themselves. This distal food provenance is critical in order for them to live in such close proximity, as observed in this study.

Approximately 90% of all modern marine sponges are demosponges, and they live in all environments from subtidal to abyssal (Rigby, 1983). Demosponges are also the most common reef sponges, where assemblages are diverse, and they can live in very high densities. Hypercalcified demosponges (external skeleton of carbonate in addition to or instead of silica) are almost the only group of Permian demosponges to have genera survive the end-Permian extinction (Finks, 2010). The only other group to survive is hypercalcified Calcarea, and no non-hypercalcified demosponges persisted (Finks, 2010). Only two or three non-hypercalcified sponge families survived. The tough rigid carbonate exoskeleton is thought to strengthen sponges for inhabiting a reef environment, and is produced by photosynthetic symbiotic cyanobacteria.

The cylindrical sponges in the Rex Chert at Ski Lake are surrounded by a halo of coarse carbonate spar, which could be a recrystallized remnant of some sort of calcareous skeleton surrounding the central spiculitic sponge. These sponges are hosted in a carbonate-cemented fine to medium sandstone, representative of a shallow water, near-shore environment. Sponges living in such an environment would need to be rigid to withstand wave action, so perhaps they are hypercalcified demosponges.

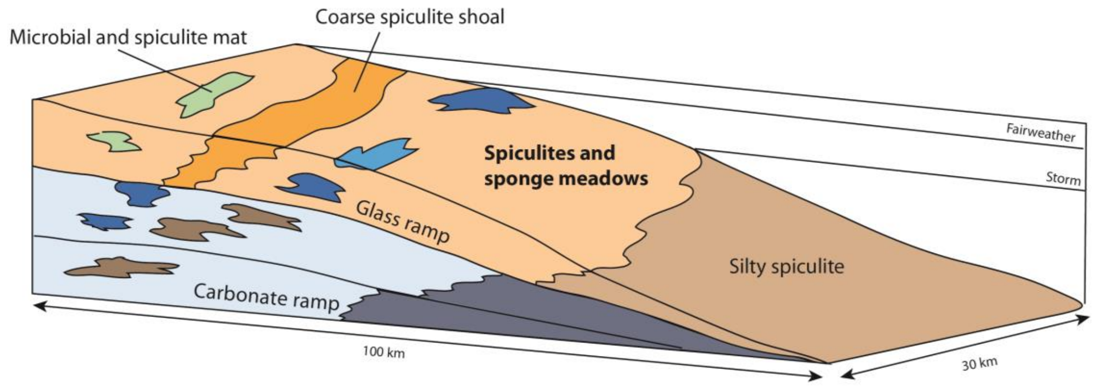


Figure 19. Generalized glass ramp model showing the occurrence of a glass ramp (tan) overtop a carbonate ramp (blue). Sponge meadows are the glass factory that produces the biosiliceous sediment of the glass ramp. Blue patches indicate small accumulations of carbonate organisms.

Glass Ramp Model

Suppression of carbonate production was probably one of the most important factors for accumulation of thick spiculites (Murchey, 2004). This interruption of carbonate production would have occurred a few times throughout the middle to late Permian, represented by the three chert members of the Phosphoria Formation.

Occurrence of rare linguloid and orbiculoid brachiopods in spiculitic chert, as well as the occurrence of mud cracks just below chert beds, suggests no deeper than moderate water depths for sponge faunas (Yocheson, 1968). Yochelson (1968) also discusses thick lenses of bioclastic limestone within the Rex Chert, and cites a number of hypotheses attempting to explain their occurrence. Bounded above and below by chert, these limestones indicate that there is a close depositional relationship between the two facies. Because spiculites occur in close stratigraphic association with clastic carbonate throughout the Phosphoria Basin and elsewhere, an alternative mode of biomineralized sediment accumulation is proposed for the Phosphoria sea: a glass ramp (Gates et al., 2004). A glass ramp is a coastal deposit dominated by material from siliceous sponges, and occupies a similar tectonic setting to carbonate ramps on continental margins (Gates et al., 2004). In deposits in Svalbard for example, there are hundreds of meters of spiculitic chert interbedded with coarse grained bioclastic carbonate, indicating deposition in a glass ramp system that occupied a similar environment to a carbonate ramp (Blomier et al., 2013; Matysik et al., 2017).

The thick spiculitic chert deposits of the Phosphoria Formation are herein interpreted to represent a glass ramp depositional system that was present for much of the span of Phosphoria deposition. This glass ramp was a dominant depositional environment at least twice in Permian northwest Wyoming, represented by the thick spicule-rich Rex and Tosi chert members. The basal Lower chert member is restricted only to a narrow north-south trending zone near the Idaho-Wyoming border, and likely represents the first and more isolated occurrence of a glass ramp.

In the sections logged for this study, the sponge zones and thick chert intervals are often adjacent to bioclastic sandstone and dolomite. The glass ramp would have mostly overtaken distal parts of the existing carbonate ramp. The lateral extent and density of sponges in the Gros Ventre Range suggest that they out-paced sedimentation by most carbonate biomineralizers, making the sea floor uninhabitable to them. The resulting bedded chert contains intervals of spectacularly preserved sponge meadows, which were the ‘glass factory’ that produced the high volumes of spicules found throughout the stratigraphy, and resulted in thick spiculitic chert.

In the Rex Chert of the Gros Ventre Range, the sponges are preserved in a matrix of sandstone or mixed sand and carbonate. Spicules are common sedimentary components in these sandstones, but spicules are not the dominant sediment grains, so a glass ramp was probably hindered by an influx of sand in northwest Wyoming. Still, sponges were living in these sandy depositional settings, and conditions allowed for them to be preserved in life position. There is evidence in the thick, spicule-rich Rex Chert in the central part of the basin that large volumes of sponges existed, and may represent a much larger glass ramp system in the deeper basin than the Tosi Chert. The Tosi Chert is mostly present and thickest in Wyoming and Montana, and suggests a larger marine transgression than the Rex Chert. On a broader scale, other thick Permian spiculitic chert deposits suggest that the glass ramp model applies to a much broader region than just the Phosphoria basin.

Most chert in the rock record, including the dark, bedded chert of the Phosphoria Formation has been long interpreted as deep-water siliceous ooze sedimentation, without a good explanation for its precipitation and accumulation. The glass ramp model improves on our understanding of Phosphoria chert accumulation in two ways: it provides a mechanism for the thick accumulation of biosiliceous sediment, and it suggests that the chert is formed in shallow water.

Permian Chert Event

Evidence for large volumes of sponges in the Permian extends beyond the Phosphoria Basin, south to central Nevada (Murchey, 2004), and north to western Canada, Alaska, Arctic Canada, and Svalbard (Fig. 20), where much thicker (hundreds of m) spiculitic chert deposits of the same age are common. These deposits as a whole are called the Permian spiculite belt (Murchey, 2004), and represent the Permian Chert Event (PCE), a widespread deposition of thick, bedded chert. The thickest spiculite deposits formed in epicratonic basins, likely during highstands. Deeper-water environments were increasingly dominated by radiolarians. The middle-late Permian was a time when biogenic silica was produced at a faster rate, or was better-preserved than at other times in earth history. Likewise, there are times in earth history when little if any biogenic chert is produced, such as the Early Triassic ‘chert gap’ (Beauchamp and Baud, 2002).

From the Carboniferous to the Sakmarian (early Permian), biogenic silica was produced along the western margin of Pangea, depositing in deep-water distal slope to basinal settings, adjacent to carbonate factories. Silica production however, was generally not high enough to result in bedded chert. Instead, cherty carbonate or shale was deposited (Beauchamp and Baud, 2002). The onset of large-scale biogenic silica production is best constrained from the Sverdrup basin in Norway, and occurred in the late Sakmarian to early Artinskian (Beauchamp and Baud, 2002). There are over 1 km of cherts in the Sverdrup basin deposited between that time and the end-Permian, representing the main part of the PCE. Evidence for landward expansion of silica factories and contraction of carbonate factories at this time can be found throughout the entire western Pangean margin. Much of these deposits are deep water radiolarian-dominated cherts, deposited adjacent to seamounts and island arcs, and are now exposed in accreted terranes from California to Alaska. The Rex and Tosi Members of the Phosphoria Formation are examples of shallower-water, shelf to ramp spicule-dominated chert deposited along the continental margin (Beauchamp and Baud, 2002). The Lopingian was the time of peak biogenic silica production along the spiculite belt, though Phosphoria Formation deposition likely ceased in the middle Lopingian.

Sponge body fossils are not common in the thick cherts outside of the Phosphoria basin, or have not been recognized. Interestingly, Matysik et al. (2017) describe what they interpret as fluid escape structures, which are “vertical pipes (<15 cm across and less than a meter high) composed of densely packed, expelled spicules” within the thick cherts of the Sverdrup Basin in Svalbard. This description sounds quite similar to the cylindrical chert interpreted herein as sponge body fossils. Fluid escape structures were one past interpretation of the Phosphoria cylinders, but most workers ruled it out as a sufficient mechanism, myself included. I would suggest that we revisit these spicule-dense cylindrical structures in chert deposits worldwide, and consider the possibility that they are actually sponge body fossils preserved as solid chert.

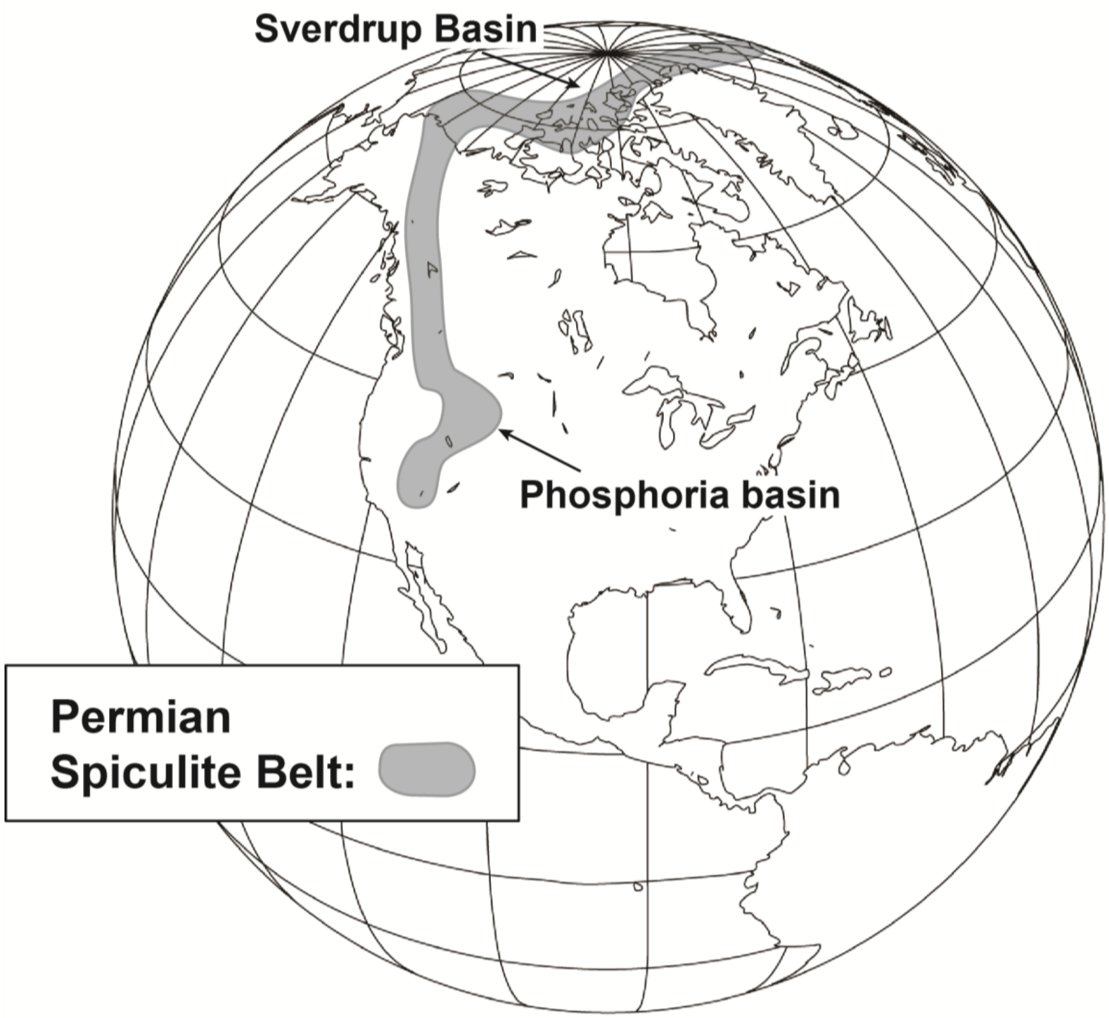


Figure 20. Map of modern distribution of Permian Spiculite Belt deposits in North America and Arctic Norway. After Murchey and Jones (1994), Beauchamp and Baud (2002), Gates et al. (2004), Murchey (2004).

CONCLUSIONS

1. Four stratigraphic sections were measured and sampled in the chert-rich Permian Phosphoria Formation in Northwest Wyoming, revealing the need for an improved understanding of the accumulation of thick, poorly studied, bedded, spiculitic chert units.
2. This is the first study to provide sufficient evidence to suggest that cylindrical chert concretions that occur in both the Rex and Tosi Chert Members in northwest Wyoming are body fossils of siliceous sponges, and not *Skolithos grandis* burrows.
3. This study proposes a glass ramp depositional model for the Rex and Tosi Chert Members, periodically hindering carbonate production on the lower ramp within the Phosphoria Basin. This model better explains observations of these and other chert units, and suggests that the thick, bedded spiculitic cherts are formed in shallow water, rather than a deep basinal setting.
4. The chert-rich Phosphoria Formation is coincident with the Permian Chert Event, and represents the southernmost extent of the Permian spiculite belt, one most extensive chert deposits in earth history. This event immediately precedes the end-Permian mass extinction event, so the cherts of the Phosphoria Formation may be a useful record of changing oceanographic conditions in the middle-late Permian.

REFERENCES

Ahr, W.M., 1973, Carbonate ramp; alternative to the shelf model: Transactions, Gulf Coast Association of Geological Societies, v. 23, p. 221-225.

Andersson, K., and Sauvagnat, J., 1998, Large columnar structures hiding Skolithos grandis n. sp. In Permian strata western U.S.A.: Ichnos, v. 6, no. 3, p. 167-182.

Beauchamp, B., and Baud, A., 2002, Growth and demise of Permian biogenic chert along northwest Pangea: evidence for end-Permian collapse of thermohaline circulation: paleogeography, Paleoclimatology, Paleoecology, v. 184, p. 37-63.

Bentor, Y.K., 1980, Phosphorites – the unsolved problem, *in*, Bentor, Y.K., ed., Marine Phosphorites – Geochemistry, Occurrence, and Genesis: Society of Economic Paleontologists and Mineralogists Special Publication 29, Tulsa, OK, p. 3-18.

Blackwelder, E., 1916, A reconnaissance of the phosphate deposits in western Wyoming: U.S. Geological Survey Contributions to Economic Geology, v. 470, p. 452-483.

Blomeier, D., Dustira, A.M., Forke, H., Scheibner, C., 2013, Facies analysis and depositional environmentsof a storm-dominated, temperate to cold, mixed siliceous–carbonate ramp: the Permian Kapp Starostin Formation in NE Svalbard: Norwegian Journal of Geology v. 93, p. 75-94.

Boury-Esnault, N., Rützler, K. (Eds.), 1997, Thesaurus of sponge morphology: Smithsonian Contributions to Zoology, no. 596, p. 1-54.

Bromley, R., Curran, H., Frey, R., Gutschick, R., and Suttner, L., 1975, Problems in interpreting unusually large burrows, *in*, Frey, R., ed., The Study of Trace Fossils: Springer-Verlag, New York, p. 351-376.

Carroll, A., Stephens, N., Hendrix, M., and Glenn, C., 1998, Eolian-derived siltstone in the Upper Permian Phosphoria Formation: Implications for marine upwelling: Geology, v. 26, no. 11, p. 1023-1026.

Cheney, T. M., Sheldon, R. P., Waring, R. G., and Warner, M. A., 1951, Stratigraphic sections of the Phosphoria Formation in Wyoming, 1951: U.S. Geological Survey Trace Elements Investigations Report 337, p. 1-28.

Cressman, E., and Swanson, R., 1964, Stratigraphy and petrology of the Permian rocks of southwestern Montana: U.S. Geologic Survey Professional Paper 313-C, p. 275-569.

Davydov, V., Crowley, J., Schmitz, M., and Snyder, W., 2016, New U–Pb constraints identify the end-Guadalupian and possibly end-Lopingian extinction events conceivably preserved in the passive margin of North America: implications for regional tectonics: Geologic Magazine, p. 1-13.

Finks, R.M., 2010, Hypercalcified demosponges and the end-Permian extinction: Global and Planetary Change, v. 73, no. 1-2, p. 141-148.

Finks, R.M., Yochelson, E.L., Sheldon, R.P., 1961, Stratigraphic implications of a Permian sponge occurrence in the Park City Formation of western Wyoming: Journal of Paleontology, v. 35, p. 564-568.

Gableman, J., 1955, Cylindrical structures in Permian(?) siltstone, Eagle County, Colorado: Journal of Geology, v. 63, p. 214-227.

Gates, L.M., James, N.P., and Beauchamp, B., 2004, A glass ramp: shallow-water Permian spiculitic chert sedimentation, Sverdrup Basin, Arctic Canada: Sedimentary Geology, v. 168, p. 125-147.

Hein, J. R., Perkins, R. B., and McIntyre, B. R., 2004, Evolution of thought concerning the origin of the Phosphoria Formation, Western US Phosphate Field, *in* Hein, J. R., ed., Life Cycle of the Phosphoria Formation: Elsevier, Amsterdam, Netherlands, v. 8, p. 19-42.

Hiatt, E.E., and Budd, D. A., 2001 Sedimentary phosphate formation in warm shallow waters: new insights into the paleoceanography of the Permian Phosphoria Sea from analysis of phosphate oxygen isotopes: Sedimentary Geology, v. 145, p. 119-133.

Jasinski S., Lee, W., and Causey, J., 2004, The history of production of the Western Phosphate Field, *in* Hein, J. R., ed., Life Cycle of the Phosphoria Formation: Elsevier, Amsterdam, Netherlands, v. 8, p. 45-71.

Kulas, H. A., and Batten, R. L., 1997, Silicified gastropods from the Permian Phosphoria rock complex of Wyoming: Contributions to Geology, University of Wyoming, v. 31, no. 2, p. 33-58.

Maldonado, M., Aguilar, R., Bannister, R., Bell, J., Conway, K., Dayton, P., Diaz, C., Gutt, J., Kelly, M., Kenchington, E., Leys, S., Pomponi, S., Rapp, H., Rutzler, K., Tendal, O., Vacelet, J., and Young, C., Sponge grounds as key marine habitats: a synthetic review of types, structure, functional roles, and conservation concerns, *in*, Rossi, S., ed., Marine Animal Forests, Springer International Publishing, Switzerland.

Maliva, R.G., Knoll, A.H., and Siever, R., 1989, Secular change in chert distribution: a reflection of evolving biological participation in the silica cycle: Palaios, v. 4, no. 6, p. 519-532.

Maliva, R.G., and Siever, R., 1989, Nodular chert formation in carbonate rocks: The Journal of Geology, v. 97, no. 4, p. 421-433.

Matysik, M., Stemmerik, L., Olaussen, S., and Brunstad, H., 2017, Diagenesis of spiculites and carbonates in a Permian temperate ramp succession – Tempelfjorden Group, Spitsbergen, Arctic Norway: Sedimentology, v. 56, no. 3, p. 745-774.

Maughan, E.K., 1984, Geological setting and some geochemistry of oil shales in the Permian Phosphoria Formation, *in*, Woodward, J., Meissner, F.F., and Clayton, J.L., eds., Hydrocarbon Source Rocks of the Greater Rocky Mountain Region: Rocky Mountain Association of Geologists, Denver, CO, p. 281-294.

McKelvey, V. E., 1946, Preliminary report on stratigraphy of the phosphatic shale member of the Phosphoria Formation in western Wyoming, southeastern Idaho, and northern Utah: U.S. Geological Survey Open-File Report 47-2, 162 p.

McKelvey, V.E., Sections, W., Steele Williams, J., Sheldon, R.P., Cressman, E.R., Cheney, T.M., and Swanson, R.W., 1959, The Phosphoria, Park City and Shedhorn Formations in the Western Phosphate Field: Geology of Permian Rocks in the Western Phosphate Field, Geological Survey Professional Paper 313-A, 47 p.

Murchey, B.L., 1990, Age and depositional setting of siliceous sediments in the upper Paleozoic Havallah sequence near Battle Mountain, Nevada: implications for the paleogeography and structural evolution of the western margin of North America, *in*, Harwood, D.S., and Miller, M., eds., Paleozoic and Early Mesozoic Paleogeographic Relations: Sierra Nevada, Klamath Mountains, and Related Terranes: Geologic Society of America Special Publication, v. 255, p. 137-155.

Murchey, B.L., 2004, Regional analysis of spiculite faunas in the Permian Phosphoria basin: implications for paleoceanography, *in*, Hein, J. R., ed., Life Cycle of the Phosphoria Formation: Elsevier, Amsterdam, Netherlands, v. 8, p. 111-135.

Murchey, B.L., and Jones, D.L., 1992, A mid-Permian chert event: widespread deposition of biogenic siliceous sediments in coastal, island arc and oceanic basins: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 96, p. 161-174.

Peterson, J.A., 1984, Permian stratigraphy, sedimentary facies, and petroleum geology, Wyoming and adjacent area, *in*, Goolsby, J., Motron, D., eds., The Permian and Pennsylvanian Geology of Wyoming: Wyoming Geological Association Guidebook, v. 35, p. 25-64.

Peterson, J. A., Gosman, R. F., and Swanson, R. W., 1953, Stratigraphic sections of the Phosphoria Formation in Montana, 1951: U.S. Geological Survey Circular 326, 27 p.

Piper, D. Z., and Link, P. K., 2002, An upwelling model for the Phosphoria Sea: A Permian, ocean-margin sea in the northwest United States: American Association of Petroleum Geology Bulletin, v. 86, no. 7, p. 1217-1235.

Pommer, M., 2017, Paleoenvironmental dynamics in the middle Permian Phosphoria seaway: a record from isotopic and biochemical trends, Rocky Mountain region, USA: Geological Society of America, Abstracts with Programs, v. 49, no. 6.

Pommer, M., and Sarg, R., 2017, Chemical diagenesis and biota in stratigraphic context: the Phosphoria rock complex (Permian), Rocky Mountain Region, USA: American Association of Petroleum Geologists, Annual Convention and Exhibition Abstracts with Programs, Houston, TX.

Richards, R., and Mansfield, G., 1912, The Bannock overthrust: a major fault in southeastern Idaho and northeastern Utah: Journal of Geology, v. 20, p. 683-689.

Rigby, J.K., 1983, Fossil Demospongia: Notes for a Short Course, Studies in Geology, v. 7, p. 12-39.

Rigby, J.K., and Boyd, D.W., 2004, Sponges from the Park City Formation (Permian) of Wyoming: Journal of Paleontology, v. 78, no. 1, p. 71-76.

Rigby, J.K., and Senowbari-Daryan, B., 1996, Gigantospongia, new genus, the largest known Permian Sponge, Capitan limestone, Guadalupe Mountains, New Mexico: Journal of Paleontology, v. 70, no. 3, p. 347-355.

Royse, F., 1993, An overview of the geologic structure of the thrust belt in Wyoming, northern Utah, and eastern Idaho, *in*, Snoke, A.W., Steidtmann, J.R., and Roberts, S.M., eds., Geology of Wyoming, Memoir, Geological Survey of Wyoming, Laramie, WY, v. 5, p. 272-311.

Scotford, D., and Knecht, M., 1990, Bent chert tubes in the Phosphoria Formation of Wyoming: pressure solution strain markers: Geology, v. 18, p. 846-849.

Sheldon, R. P., 1957, Physical stratigraphy of the Phosphoria formation in northwest Wyoming: U.S. Geological Survey Bulletin, 1042-E, 185 p.

Sheldon, R. P., 1963, Physical stratigraphy and mineral resources of Permian rocks in western Wyoming: U.S. Geological Survey, Professional Paper 313-B, p. 49-273.

Sheldon, R.P., 1984, Polar glacial control on sedimentation of Permian phosphorites of the Rocky Mountains, USA, *in*, Proceedings of the 27th International Geological Congress, Moscow, p. 223-243.

Sheldon, R. P., Waring, R. G., Warner, M. A., and Smart, R. A., 1963, Stratigraphic sections of the Phosphoria Formation in Wyoming 1949-50: U.S. Geologic Survey Survey 307, p.

Stephens, N., and Carroll, A., 1999, Salinity stratification in the Permian Phosphoria sea; a proposed paleooceanographic model: Geology, v. 27, no. 10, p 899-902.

Tapanila, L., and Knecht, R.J., 2009, A Carboniferous sponge forest on the carbonate ramp of the Antler foreland basin, central Idaho: Geological Society of America, Abstracts with Programs, v. 41, no. 6, p. 15.

Tapanila, L., Pruitt, J., Pradel, A., Wilga, C., Ramsay, J., Schalder, R., and Didier, D., 2013, Jaws for a spiral-tooth whorl: CT images reveal novel adaptation and phylogeny in fossil Helicoprion: Biology Letters, v. 9, no. 2, p. 1-4.

Thornburg, J.M.B., 1990 Petrography and sedimentology of a phosphatic shelf deposit: the Permian Shedhorn Sandstone and associated rocks in southwest Montana and northwest Yellowstone National Park: PhD Dissertation, University of Colorado, Boulder, 719 p.

Wardlaw, B., 2015, Gondolellid conodonts and depositional setting of the Phosphoria Formation: Micropaleontology, v. 61, no. 4-5, p. 335-368.

Wardlaw, B. R., 1977, The biostratigraphy and paleoecology of the Gerster Formation (Upper Permian) in Nevada and Utah: U.S. Geological Survey, Open File Report 77-470, 125 p.

Wardlaw, B. R., 1979, Transgression of the Retort Phosphatic Shale Member of the Phosphoria Formation (Permian) in Idaho, Montana, Utah, and Wyoming: US Geologic Survey Professional Paper 1163-A pp. 1-4.

Wardlaw, B. R., and J. W. Collinson, 1986, Paleontology and deposition of the Phosphoria Formation: Contributions to Geology, University of Wyoming, v. 24, p. 107-142.

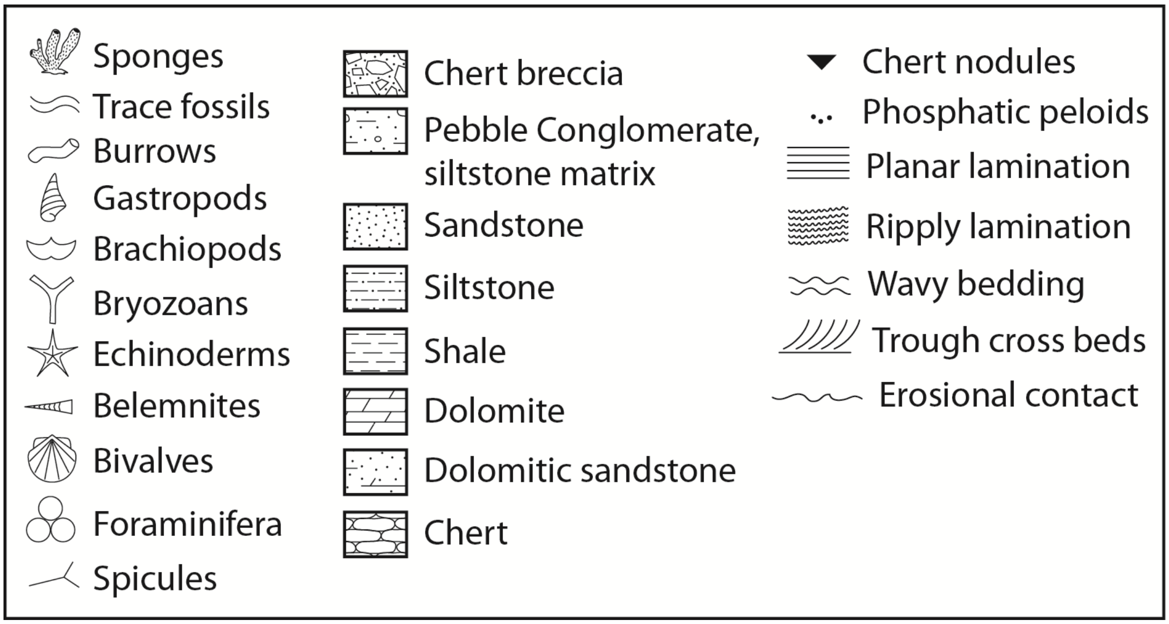
Whalen, M.T., 1996, Facies architecture of the Permian Perk City Formation, Utah and Wyoming: implications for the paleogeography and oceanographic setting of western Pangea, *in* Longman, M.W., and Sonnenfeld, eds., Paleozoic systems of the Rocky Mountain Region: Society for sedimentary geology, Denver, CO p. 355-378.

Yochelson, E. L., 1968, Biostratigraphy of the Phosphoria, Park City, and Shedhorn Formations: U.S. Geological Survey, Professional Paper 313-D, p. 571-660.

APPENDICES

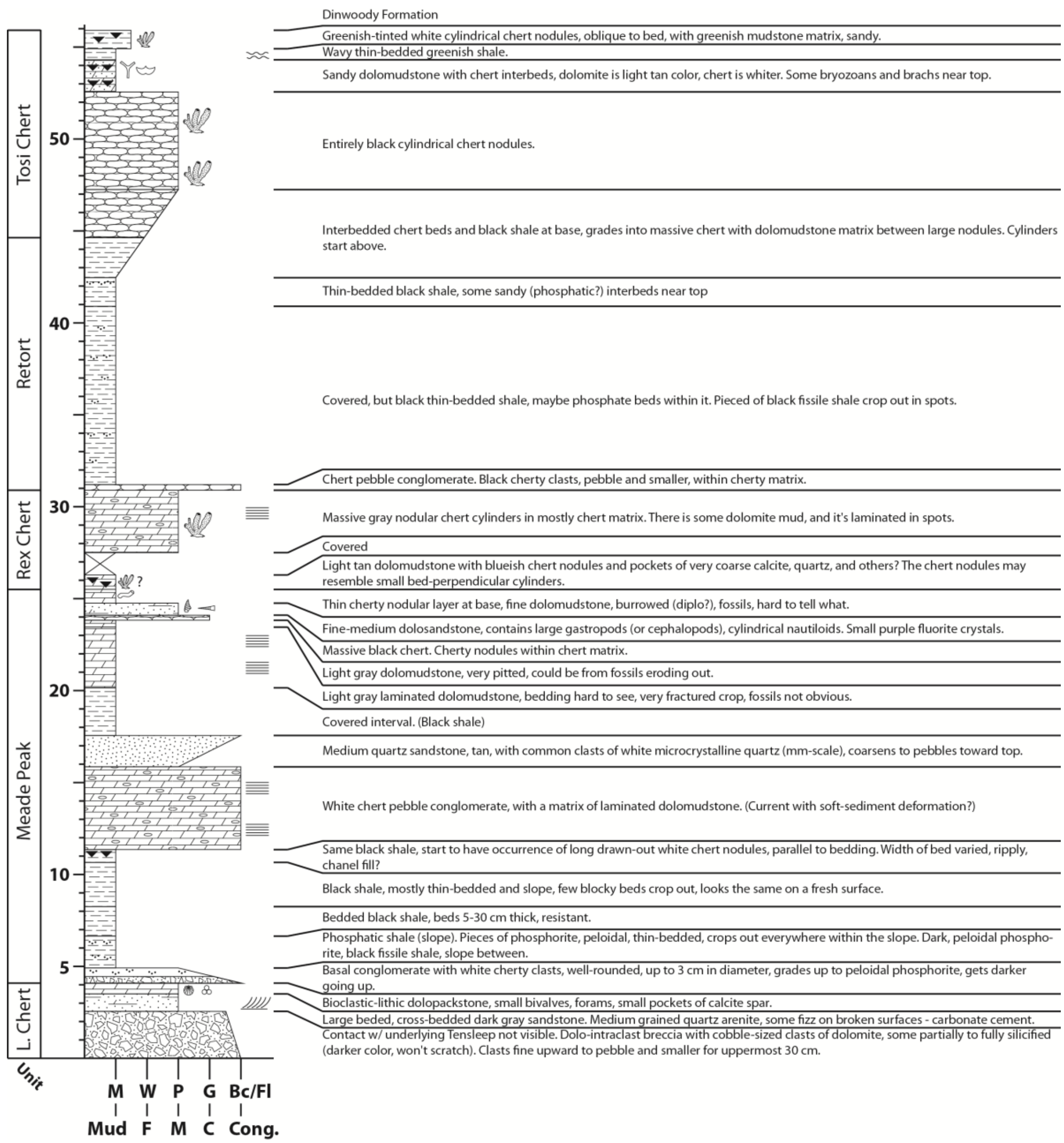
Appendix A: Stratigraphic Columns with Descriptions

Key to Symbols



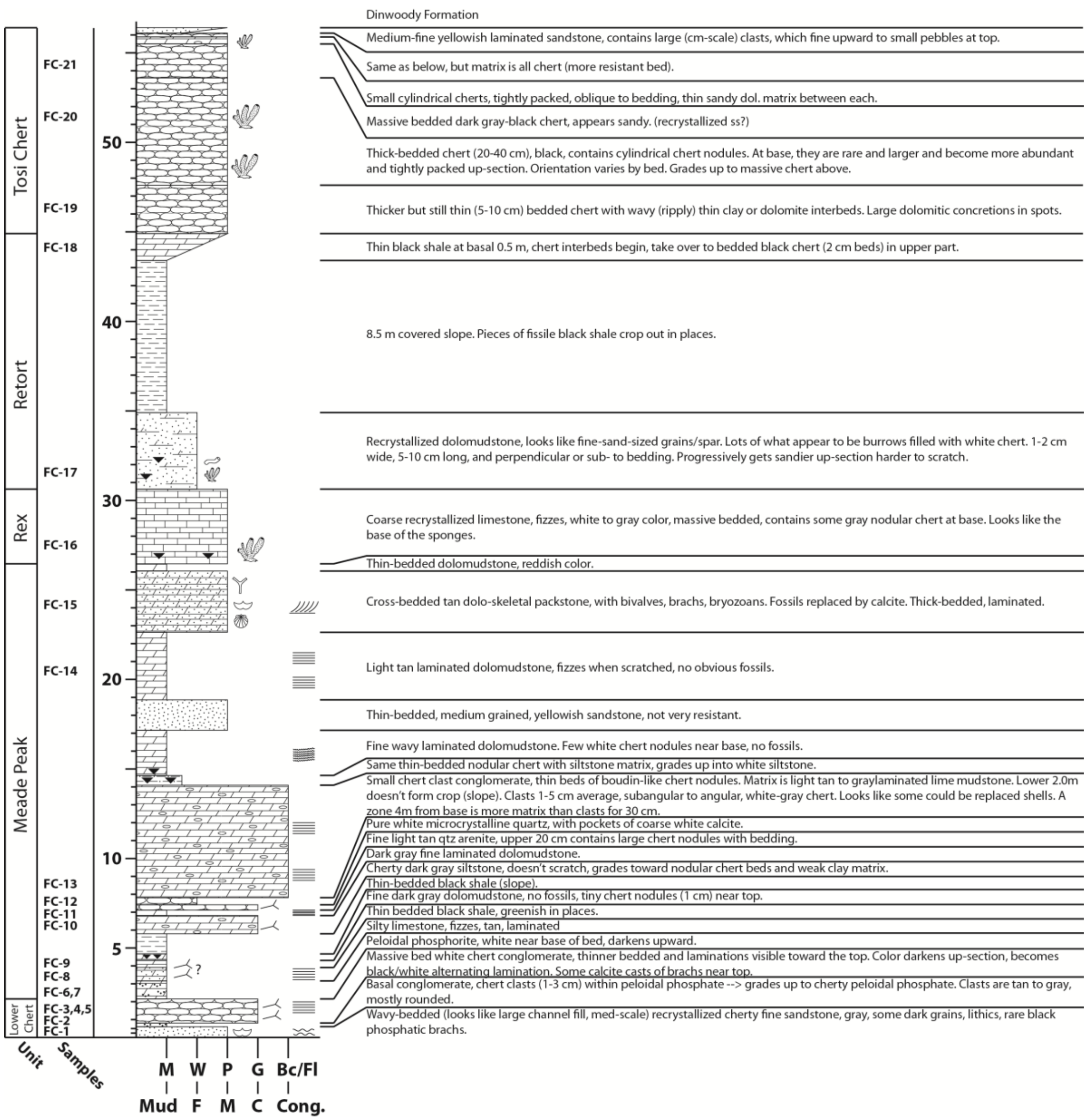
Crystal Creek, Wyoming
UTM 12T 0533289 4821386
Elevation 7548 ft.
Strike and Dip 356, 13 (E)

Drive east on Gros Ventre Road, from Kelly, Wyoming. After about 13 miles, just past the Crystal Creek Campground, turn right onto Forest Road 30377, heading south. Park at wilderness boundary trailhead at the end of the road. Phosphoria is extensively exposed in a cliff band on the east side of Crystal Creek, starting just south of the parking area.



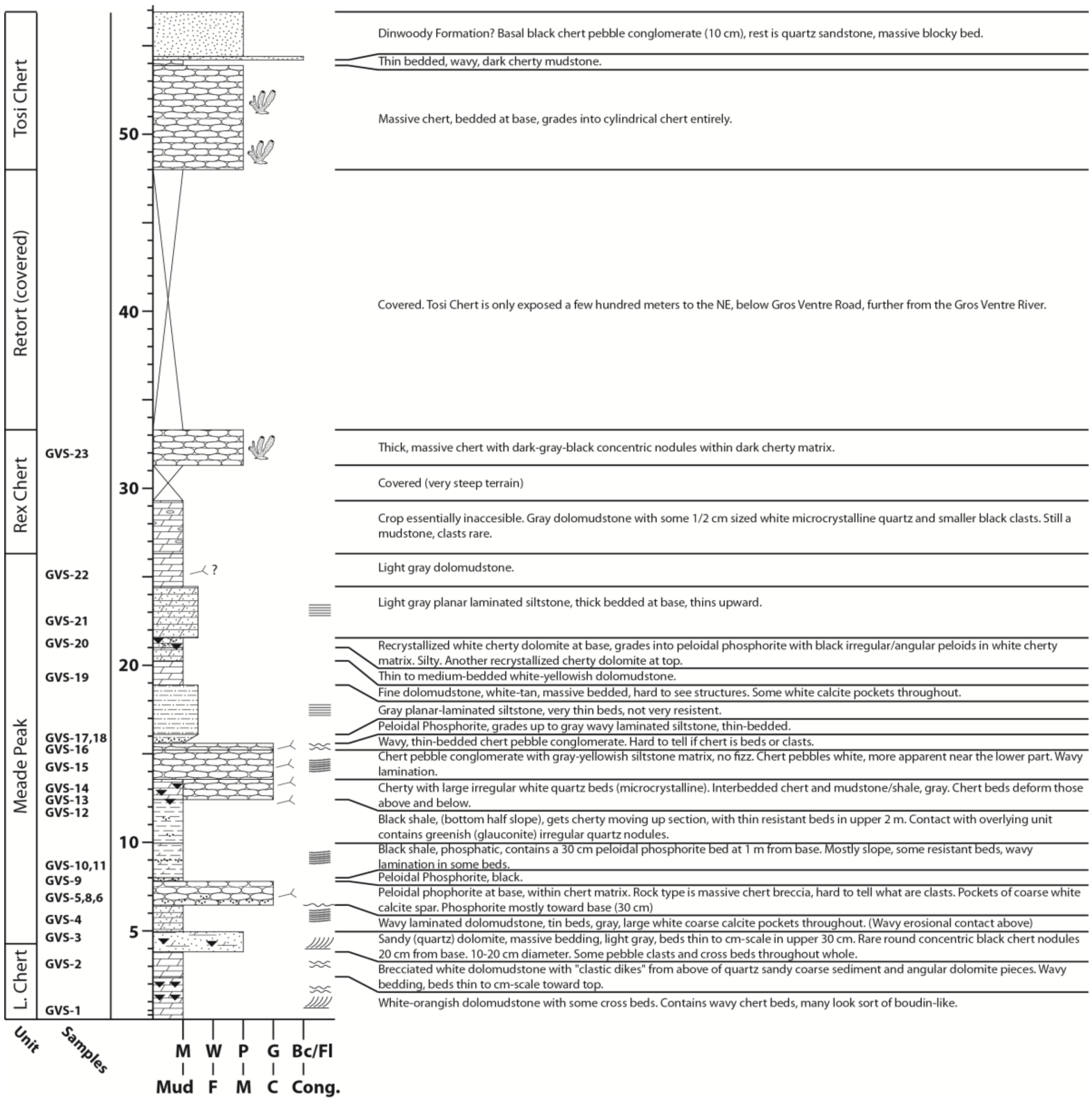
Flat Creek, Wyoming
UTM 12T 0533285 4821385
Elevation 7208 ft.
Strike and Dip 290, 22 (N)

From Jackson, Wyoming, take National Elk Refuge Road heading north. After 4.8 miles, bear right onto Flat Creek Road, and drive 4.3 miles. Stay straight onto Forest Road 30442 for 2 miles, and the outcrop is on the left side of the road, a steep hike up. Exposure is continuous for at least one mile along the road.



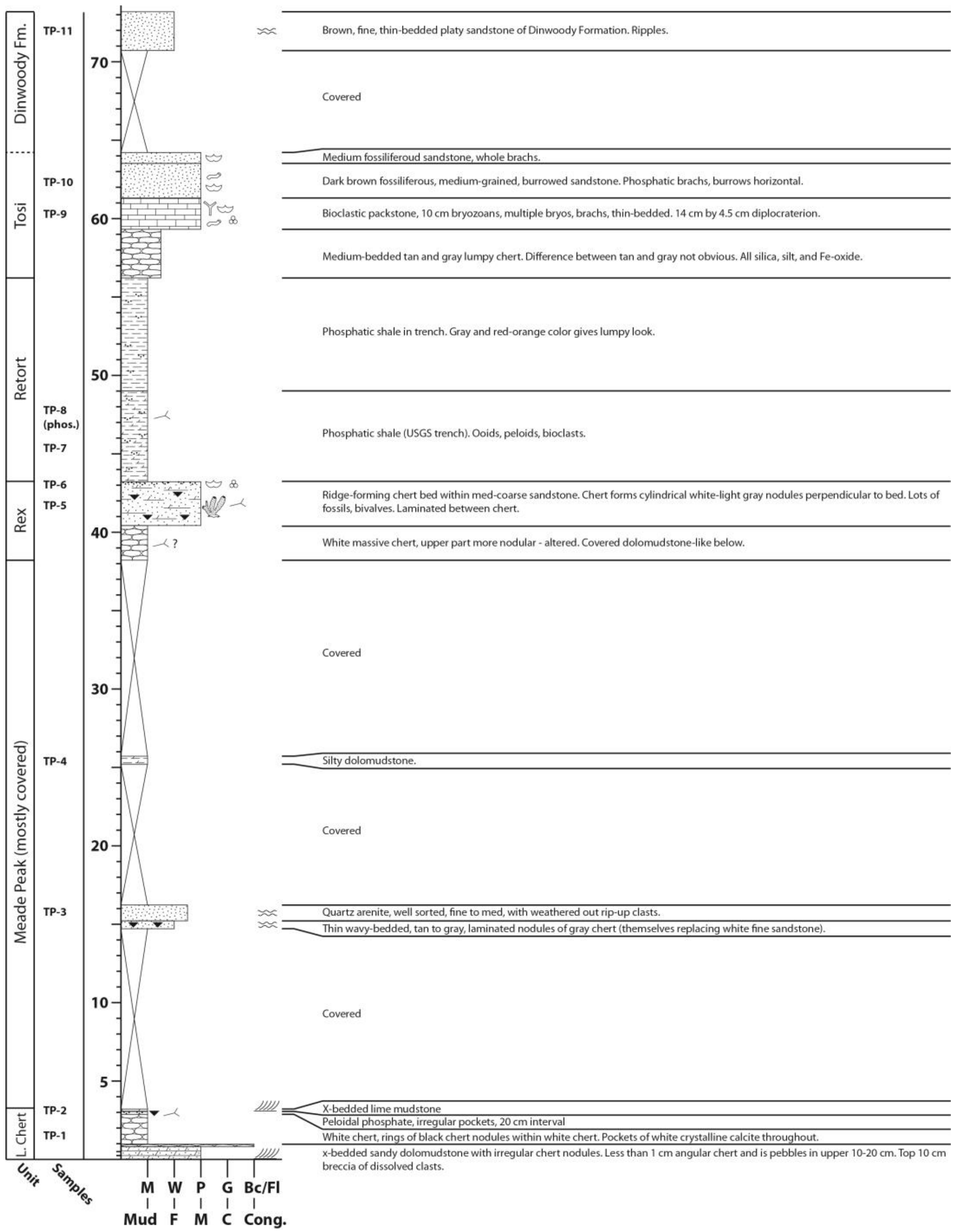
Gros Ventre Slide, Wyoming
UTM 12T 0535507 4830971
Elevation 6878 ft.
Strike and Dip 297, 23 (N)

Turn onto Gros Ventre Road north of Kelly, Wyoming, heading east. After 3.6 miles, turn right onto Forest Road 30359. Outcrop is located at the bottom of the road, just above a small pond, on the north edge of the Gros Ventre River, downstream from Lower Slide Lake.



Ski Lake, Wyoming
UTM 12T 0504538 4819965
Elevation 9509 ft.
Strike and Dip 240, 25 (NW)

Turn north onto Forest Road 30972, about 1.7 miles below the top of Teton Pass Highway, on the Jackson, Wyoming side. Park, and hike on the Phillips Canyon Trail, past Ski Lake. The trial climbs steeply up a ridge about 1 mile after Ski Lake, and Phosphoria outcrop is poorly exposed on the right side of the trail.



Appendix B: Suggestions for Future Studies

1. A future study focused on Phosphoria Formation sponges should include an analysis of the chert cylinders in Montana that are preserved as sandstone, rather than spiculitic chert. This study only focused on the sponges in northwest Wyoming, but it is very likely that they are related structures to the sandstone cylinders further north in the Phosphoria Basin. Without an analysis of outcrops, hand samples, and thin sections, it is not possible to determine if these structures are sponges as well.
2. In order to narrow down the taxonomy of the Phosphoria sponges, a biomarker analysis should be done on any trace amounts of organic matter preserved in the chert cylinders. It is very likely that they are demosponges, but no taxonomy has been worked out. Internal morphology is not well preserved, but if there is sufficient organic matter, a biomarker analysis may be possible.
3. A student should digitize stratigraphic columns for all of the 100+ stratigraphic sections logged by the USGS and published in their extensive surveys of the Phosphoria Formation. The logged sections hold a valuable amount of information, which is not well-presented in the table form that they were originally published in. Also, many of these sections were logged in trenches dug in the middle of the 20th century, and access to the same areas today is limited. Once digitized, this huge array of data could be integral in determining the type of basin that the Phosphoria represents, and would provide an excess of points through which cross-sections could be constructed to transect the basin. Currently, only a few such cross-sections exist, and many use out-dated unit names and are not drawn to scale, making the basinwide extent of members difficult to visualize.

4. Studies of the broader Permian spiculite belt are largely lacking fossil sponges responsible for producing the biosiliceous sediment preserved as chert. Some of these studies talk about spicule-filled burrows, or columnar structures in the chert, which based on the interpretations in this study, may be heavily silicified sponges. In general, more studies are needed to understand these thick middle-late Permian spiculites.

Appendix C: Petrographic Notes

Flat Creek (FC)

FC-1: Fine-medium quartz arenite, with some patchy chert. Mostly quartz overgrowth cement, some fine carbonate spar between grains. Rare larger grains look like they may be fossils replaced by chert. Very rare fish scales (?), finely laminated/fibrous microcrystalline quartz.

FC-2: Phosphorite composed of finely laminated/fibrous fish scales (?) parallel to bedding. Rare rip-ups of spiculitic chert and peloids that are yellowish in regular light. In x-polars, most grains go dark. Late stage vugs and veins filled with carbonate.

FC-3: Fine sandstone to siltstone, rather poorly sorted, carbonate cemented. It looks like initial quartz overgrowth cement, then drusy quartz filling voids (chert). Much of the chert and quartz cement later cannibalized by interstitial carbonate. Upper part of slide mostly unaltered by carbonate.

FC-4: Texture preserves spiculite. Entire rock initially recrystallized as drusy quartz, then later overprinted largely by small carbonate crystals. Carbonate preferentially did not replace spicules. (spiculitic chert)

FC-5: Spiculitic chert, completely recrystallized to drusy quartz, coraser in some spots than others. Upper part of slide presered spicules better, and spicules are preserved mostly as isopachous chalcedony. Rare fine quartz sand/silt. Some hematite (?) and more commonly carbonate crystallized along young fractures.

FC-6: Peloidal (?) phosphorous with coarse void-filling poikilotopic carbonate spar cement between garins that all goes extinct together. The peloids often have some skeletal nucleus, and concentric rings surrounding it, and their shapes appear grown in place, or at least late-stage intergrown, oriented with bedding. Phosphatic ooids?

FC-7: Similar peloidal phosphorite to FC-6, with more carbonate cement. Carbonate cement looks like recrystallized micrite in spots. Peloids look in spots to be variably recrystallized to chert, or perhaps only some of the fine sand grains within laminated peloids.

FC-8: Recrystallized silty mudstone, mostly chert (drusy quartz) between silt grains. Late stage poikilotopic carbonate overprinting much of the chert, which go extinct in large waves across slide.

FC-9: Chert (drusy quartz) nearly all replaced by fine carbonate (dol). Dolomite looks to be later-stage than chert, few textures preserved.

FC-10: Silty spiculitic chert, to siltstone in spots, mostly replaced by fine crystalline carbonate. Laminated layers within slide, define zones of more or less carbonate crystallization. Common small dark brown-reddish hematite (?) filling some interstitial space. Latest stage poikolotopic calcite along young fractures.

FC-11: Recrystallized dolomudstone (?). Blotchy fenestral-like texture, and carbonate looks secondary, replacing fine chert. Overprinted texture preserved spicules in spots, and euhedral carbonate seems to have grown inward replacing them.

FC-12: Coarse crystalline silica with large pockets of coarse calcite/dol spar. Texture of quartz indicates replacement of spiculite. Spiculite initially recrystallized to coarse euhedral quartz, with less common zebraic chalcedony, then void space filled with very coarse calcite spar, which preserved no original texture.

FC-14: Silty recrystallized dolomudstone. Quartz sand, fine, rare, with some void-filing euhedral quartz, and later-stage carbonate spar, but most still preserve void space. Rare silt-sized hematite (?) grains throughout, with orangish halos surrounding them.

FC-15: Fine-medium phosphatic sandstone. Most grains quartz sand, with about 10% phosphatic (?) peloids, and what appears to be interstitial phosphate. Limpid dolomite crystals along grain boundaries throughout. Local chert in one corner, drusy quartz. Medium quartz sand mostly in pockets, laminated.

FC-17: Fine-medium sandstone with quartz overgrowth cement, about 5% sand-sized chert (drusy Quartz) grains, and 5% rounded phosphate (?) grains. Lots of interstitial poikilotopic carbonate, looks secondary. Some brown blotchy phosphate within finer carbonate between sand grains.

FC-18: almost entirely secondary fine crystalline carbonate, some pockets coarser crystals, in darker finer matrix. Overprinted texture is evident of spiculte. Rare small hematite (?) grains throughout, with reddish edges.

Gros Ventre Slide (GVS)

GVS-1: Recrystallized dolomite, porous, Fenestral? Entirely rhomb-shaped dol crystals, with pockets that go black in cross polars. Pockets may be filled with very fine silica, or phosphate?

GVS-2: well-sorted, sub-angular fine quartz sand injected into fine crystalline carbonate host rock. Looks like some large dolomite crystals in a dolomite or calcite finer crystalline matrix. Recrystallized dolomudstone with sand injected from above.

GVS-3: Carobate cemented fine to medium quartz arenite. Cement is recrystallized to coarse spar in spots. Some large quartz sand and chert clasts, microcrystalline, some partially replaced by dolomite.

GVS-4: Sandy (angular fine-med quartz) recrystallized dolomudstone. Fine crystalline dolomite some coarse sand-sized finer-grained rip up clasts near stratigraphic bottom of slide.

GVS-5: Phosphorite. Mostly composed of strange banded grains, which go very dark under cross polars. It looks siliceous, but I have a hunch toward phosphate (fish scales?). Large pocket of specular chert with gastropod and echinoderm fossils on one end, which is bounded in part by some large fossil cutting the slide in half.

GVS-6: Sandy (qtz) recrystallized spiculitic chert. Spicules replaced by microcrystalline qtz, but overprinted texture remains. It was probably about 75% spiculed on deposition. Some large voids, porosity in spots, and small brown phosphate grains.

GVS-8: Fine quartz arenite with mostly qtz overgrowth cement, but some minor calcite spar between grains. Rare chert, and maybe carbonate clasts? There are common large pockets filled with very coarse calcite spar, replacing something an original specular texture. Carbonate and spiculitic chert rip up clasts?

GVS-9: Peloidal Phosphorous, with interlocking flat-lying elongated peloids. Slide goes almost all black under x-polars. Little internal structure, sometimes appears laminated. No mud at all between grains, grains seem to have grown in place, interlocking.

GVS-10: Peloidal phosphorous, with large concentrically laminated peloids. There are distinct horizons within bedding, which look like tiny unconformities, or perhaps short-lived hardgrounds. Lots of other fossil components, but I have no clue. Fish scales? Bones?

GVS-11: Sandy recrystallized dolomudstone. Dol is fine, crystalline, with minor amounts of angular fine sand and silt throughout. Minor fine sand-sized phosphate clasts.

GVS-12: Silty dolomudstone. Some fine-sand-silt-sized dark phosphate. Dolomite is fine, crystalline. Fine quartz sand and silt partially replaced by dolomite on grain boundaries.

GVS-13: Recrystallized spiculitic chert. Most of the specular texture has been destroyed, but it is all massive chert. There is some patchy carbonate in spots, looks like it's crystallized along fractures and lamination planes. . There is green glauconite surrounding a few large pockets of chert.

GVS-14: Fine crystalline chert, with pockets of coarse solomite spar. There are some euhedral dolomite crystals in spots, but majority of dolomite is finer. Some textures under plane light look like they originally were spiculitic chert.

GVS-15: Chert throughout, coarsely crystalline in spots. Spicule casts visible, particularly within finer crystalline chert, most texture gone. Gets sandier up-section, fine angular quartz sand, with some interstitial carbonate becoming more common up-section too.

GVS-16: Finely recrystallized specular chert, texture is obvious in most places. Large very coarse pockets of calcite spar filling pockets throughout. Calcite crystallized after chert.

GVS-17: Base is fine angular quartz arenite, with minor chert clasts, and some interstitial brown-green mineral (glauconite?) moving up-section. The upper part of the slide is mostly larger (coarse sand to small pebble) rounded chert and spiculitic chert clasts within matrix of fine quartz sand. Less common fish scales (?) and very little mud between clasts. Many chert clasts stained brownish, often autobrecciated, with veins filled by crystalline silica. Some isolated stylolites, perpendicular to bedding, between two chert clasts.

GVS-19: Recrystallized laminated dolomudstone, with some sand (less than 5%). Some wavy lamination, and what look like large (pebble) rounded intraclasts slightly coarser crystallized.

GVS-20: Sandy dolomudstone, looks partially replaced by chert. Chert came first, then carbonate. Rare ghosts of spicules, and dark elongated phosphate grains, oriented flat with bedding. Perhaps molds of the cores of spicules. Fine lamination.

GVS-21: Sandy recrystallized dolomudstone, fine sand in lenses, irregular dark lamination visible with naked eye. Looks like wavy stylolites, perpendicular to bedding, with truncated grains juxtaposed against micrite. (IS THIS EVIDENCE FOR LIMESTONE DISSOLUTION AFTER BURIAL AND VERTICAL COMPACTION? Sand grains don't look dissolved. Stylolites follow their grain boundaries!)

GVS-22: Dolomudstone with lots of small linear phosphate grains (15%). About 5% fine angular quartz sand. Mostly elongated rip-up clasts of phosphate of various textures. Turbulent-looking orientations of elongated grains.

Ski Lake (TP)

TP-2: spiculitic chert, with large carbonatic echinoderm (crinoid) skeletal pieces. Common small dolomite crystals between spicules.

TP-3: Fine quartz-chert arenite. Some tiny dol crystals between sand grains. Still a quartz overgrowth cement though. All clasts subround-round.

TP-5: Inside cylinder, it is fine chert with some euhedral dol crystals. The chert under plane light looks like it is recrystallized spicules. Outside the cylinder, it is a fine quartz sandstone with some phosphate and carbonate grains. Matriz/spar has been recrystallized to be coarse calcite/dol encompassing many grains.

TP-6: fine-medium bioclastic quartz arenite. Mud between grains in spots. Lots of fossils, hard to identify. Forams, brachs?

TP-7: Cherty recrystallized dolomudstone. Top and bottom edges altered by weathering of crop.

TP-8: Fine (med-sand-sized) peloidal phosphate. Tightly packed near base, with only fine chert between grains. Moving up-section, it becomes more silty, contains some fossils, elongated phosphatic cores of spicules? Upper ¼ of slide is cherty siltstone-shale, with dispersed peloids, and rare elongated phosphatic cores of spicules.

TP-9: Bioclastic packstone, with coarse recrystallized carbonate spar cement. Some rare mud between grains. Large brachs, bryozoans, forams(?).

TP-10: fine-medium quartz arenite, with pockets of cherty matrix between grains (silt recrystallized to chert?). Some fossils, hard to tell.

Appendix D: Table of Collected Sponge Specimens

Specimen #	Locality	Stratigraphic position	Thin section?	Description
FC-T-1	Flat Creek	Tosi Chert		Chert cylinder, brecciated, 10 cm diameter
FC-T-2	Flat Creek	Tosi Chert		Chert cylinder, 4 cm diameter
FC-T-3	Flat Creek	Tosi Chert		Chert cylinder, brecciated, cone in cone
FC-T-4	Flat Creek	Tosi Chert		Chert cylinder in silicified matrix, 4 cm diameter
FC-T-5	Flat Creek	Tosi Chert		Chert cylinder, surrounded by cone-in cone, 4 cm diameter
FC-T-6	Flat Creek	Tosi Chert, top		Sandstone cylinder, wrinkly, 6 cm diameter
FC-T-7	Flat Creek	Tosi Chert		Chert cylinder, cut by fault, 4 cm diameter
FC-T-8	Flat Creek	Tosi Chert		Two small chert cylinders, 1 and 3 cm diameter
FC-T-9	Flat Creek	Tosi Chert		Cone-in cone surrounding partial chert cylinder
FC-T-10	Flat Creek	Tosi Chert		Chert cylinder base, 2 cm diameter
FC-T-11	Flat Creek	Tosi Chert		Chert cylinder, brecciated, fault sets, 10 cm diameter
FC-T-12	Flat Creek	Tosi Chert		Chert cylinder, dark cherty matrix, 7 cm diameter
FC-T-13	Flat Creek	Tosi Chert, top		Sandy chert cylinder, 7 cm diameter
FC-T-14	Flat Creek	Tosi Chert		Chert cylinder base
FC-T-15	Flat Creek	Tosi Chert		Chert cylinder base
FC-T-16	Flat Creek	Tosi Chert		Chert cylinder, slicken lines on surface, 3 cm diameter
FC-T-17	Flat Creek	Tosi Chert, base		Spherical chert nodule, contains crinoids
FC-T-18	Flat Creek	Tosi Chert, lower		Lumpy white chert, stromatolite or sponge?
FC-T-19	Flat Creek	Tosi Chert, lower		Lumpy white chert, stromatolite or sponge?
FC-T-20	Flat Creek	Tosi Chert, base		Spherical chert nodule, contains crinoids
GVS-23	Gros Ventre Slide	Rex Chert, 32 m		Four chert cylinders in fine sandstone matrix
GVSf-1	Gros Ventre Slide	Tosi Chert	Yes	Two or more chert cylinders, cone in cone, breccia
GVS-T-1	Gros Ventre Slide	Tosi Chert		Chert cylinder, bent, 7 cm diameter
GVS-T-2	Gros Ventre Slide	Tosi Chert		Two chert cylinders, ~4 cm diameter
GVS-T-3	Gros Ventre Slide	Tosi Chert		At lest three chert cylinders, breccia, muddy matrix
GVS-T-4	Gros Ventre Slide	Tosi Chert		Chert cylinder, irregular
GVS-T-5	Gros Ventre Slide	Tosi Chert		At least 4 chert cylinders, breccia, bending, cherty matrix
GVS-T-6	Gros Ventre Slide	Tosi Chert		Chert cylinder piece, fossil in middle
GVS-T-7	Gros Ventre Slide	Tosi Chert		Chert cylinder, oval shape, 6 cm diameter
GVS-T-8	Gros Ventre Slide	Tosi Chert		Chert cylinder, fractured, 6 cm diameter
GVS-T-9	Gros Ventre Slide	Tosi Chert		Chert cylinder, breccia, faulted?
GVS-T-10	Gros Ventre Slide	Tosi Chert		At least 3 chert cylinders, chert breccia matrix

Specimen #	Locality	Stratigraphic position	Thin section?	Description
GVS-T-11	Gros Ventre Slide	Tosi Chert		Three chert cylinders, breccia, cone-in cone
GVS-T-12	Gros Ventre Slide	Tosi Chert		Large chert cylinder, chert breccia matrix, 10 cm diameter
GVS-T-13	Gros Ventre Slide	Tosi Chert		Two chert cylinders, chalcedony veins
GVS-T-14	Gros Ventre Slide	Tosi Chert		Chert cylinder and chert matrix, cone in cone, 7 cm diameter
TP-5	Ski Lake	Rex Chert, 42 m	Yes	Multiple chert cylinders in sandstone matrix
TP-R-1	Ski Lake	Rex Chert, 42 m		Multiple chert cylinders in sandstone matrix
TP-R-2	Ski Lake	Rex Chert, 42 m		Multiple chert cylinders in sandstone matrix
TP-R-3	Ski Lake	Rex Chert, 42 m		Multiple chert cylinders in sandstone matrix
TP-R-4	Ski Lake	Rex Chert, 42 m		Multiple chert cylinders in sandstone matrix
TP-R-5	Ski Lake	Rex Chert, 42 m		Multiple chert cylinders in sandstone matrix
TP-R-6	Ski Lake	Rex Chert, 42 m		Multiple chert cylinders in sandstone matrix

Appendix E: Additional Phosphoria Sponge Photographs

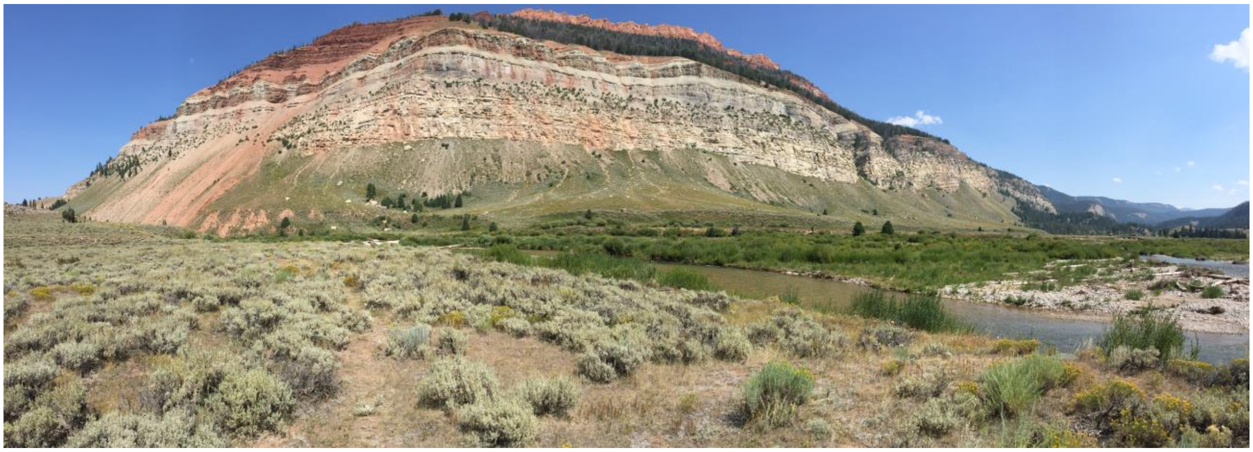


Figure A1. Panorama photograph of exposure at Crystal Creek. The Tensleep Formation is the lower cliff band, overlain by mostly slope of the Phosphoria Formation, and the overlying dark colored cliffs are the Dinwoody Formation, and younger strata. Crystal Creek logged section was near the far left side of this photograph.

Figure A2. Cylindrical chert at Crystal Creek. Bottom half of photograph is an oblique zone of chert cylinders protruding out of the outcrop. Upper half is a bedding-perpendicular zone.

Figure A3. Photograph of a well exposed zone at Crystal Creek, where adjacent cylinders are seen touching for up to 1 m. Upper half is an oblique zone, and lower half is a bedding-perpendicular zone.

Figure A4. Photograph of most of the thickness of the Tosi Chert at Crystal Creek. The different zones of oblique and bedding-perpendicular cylinders are apparent.

Figure A5. Photograph of a bedding-perpendicular cylinder zone at Crystal Creek. Cylinders are very wide, and slightly spaced out.

Figure A6. Photograph of the base of the Tosi Chert at Flat Creek. This thin-bedded black chert overlies the Retort Member, and underlies the cylindrical chert zones.

Figure A7. Photograph from the Tosi Chert at Flat Creek. Left of the rock hammer handle is the narrowed base of one cylinder. To the right may be the top of a cylinder exposed, and appears to narrow like the bases. Both of these were collected, however, they revealed no difference in internal morphology than other samples.

Figure A8. Photograph of an oblique zone of cylindrical chert in the Tosi Chert at Flat Creek. Individuals can be traced for over a meter.

Figure A9. Photograph of cylindrical chert in the Tosi Member at Flat Creek. Cylinders at the base are bedding-perpendicular, above which, they are oblique to the right, then curve back to the left for another oblique zone.

Figure A10. Photograph of the Tosi Chert at Gros Ventre Slide. Zones of different cylindrical chert orientations can be seen.

Figure A11. Photograph of the cylindrical chert zones at Gros Ventre Slide. At least two oblique zones are well exposed.