Determining the Extent of Hothouse Climate Effects on the Jurassic Silica Cycle

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Master of Science

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This thesis titled

Determining the Extent of Hothouse Climate Effects on the Jurassic Silica Cycle

by

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ABSTRACT

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A global assessment of the spatial and temporal distribution of Jurassic bedded and nodular cherts undertaken in this study reveals significant findings related to the Jurassic silica cycle. Results show that the earliest Jurassic cherts were mostly deposited in shelf and peritidal settings. This finding challenges the conventional thinking that most Jurassic cherts originated in deep water settings. The Central Atlantic Magmatic Province (CAMP) Large Igneous Province (LIP) is interpreted to be the source of sufficient silica to the oceans to generate shelf and peritidal chert deposits, just after the end-Triassic extinction. The Pleinsbachian-Toarcian HEATT (Haline Euxinic Acidic Thermal Transgression) episode caused in part by the Karoo-Ferrar LIP led to a relative increase in shelf-originated chert deposits compared to previous periods. Radiolarian cherts were more abundant than expected during HEATT conditions. The low radiolarian diversity at this time shows that some radiolarian taxa could thrive as opportunists and extract significant amounts of dissolved silica from the Early Jurassic ocean. Increased chert deposition during the Late Jurassic is coincident with three seafloor LIPs that emitted a combined basalt volume nearly equivalent to the Pliensbachian-Toarcian Karoo-Ferrar LIP. The Middle Jurassic and older parts of the Upper Jurassic (Oxfordian-Kimmeridgian) have an increase in chert deposition without significant LIP
activity. The chert deposits of Aalenian through Tithonian age might be an indication of the increasing influence of diatoms on the Jurassic silica cycle.
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CHAPTER 1: INTRODUCTION

Biogenic chert is an important component of the record of Earth’s ancient oceans. Most chert accumulation in the Phanerozoic is thought to have been strongly influenced by silica-secreting organisms that extracted dissolved silica from ancient seawater and left a record of that activity as the biosiliceous skeletal material was converted to chert via diagenesis (Maliva et al., 1989). Maliva et al. (1989) outlined a stepwise progression of styles of chert accumulation as silica-secreting organisms evolved and suggested that even though a well-known major change occurred when Phanerozoic silica secretors began to control the patterns of chert accumulation as they began to draw down the concentration of dissolved silica from higher Precambrian levels, the changes in the nature of chert accumulation did not stop after the Precambrian ended. Subsequent research (e.g., Racki and Cordey, 2000; Kidder and Erwin, 2001) built on this work and showed how other factors could affect the nature of biogenic chert accumulation. Some of these factors include volcanic pulses of new supplies of silica and nutrients and changes in changes in paleoceanography and/or paleoclimate that act in various ways to perturb the silica cycle.

Silicon is second only to oxygen in Earth’s crustal abundance (Skinner, 1979). These two elements commonly bond together to form the silica (SiO$_2$) that is a fundamental building block of silicate minerals. Despite the abundance of silica in the crust, modern oceans are undersaturated with respect to this compound (Skinner, 1979). About 95% of dissolved oceanic silicon is in the form of silicic acid (Si(OH)$_4$). Geological and biological processes determine the amount of oceanic silicic acid (Figure
1.1). Terrestrial-formed silicic acid is brought to surface waters via silicate weathering of continental crust. Rivers are the dominant mechanism for delivering silicic acid to the oceans (Tréguer et al., 1995; Aguilar-Islas and Bruland, 2006; Opfergelt et al., 2013). Considerably lesser, but significant amounts of silica are delivered as dust blown into the ocean by winds (Tréguer et al., 1995), although the degree to which this particulate silica dissolves is not well investigated to date (De La Rocha, 1996). Relative to riverine delivery, the deep ocean also receives minor amounts of silicic acid through weathering of the oceanic crust, as well as from siliceous minerals that form from hydrothermal vents.

Planktonic organisms like diatoms and radiolarians create their skeletons by using silicic acid from the ocean. Once these organisms die, the biogenic silica from their skeletons either dissolves in the seawater or settles on the ocean floor and accumulates with the oceanic sediment. Tréguer et al. (1995) pointed out that dissolution is so prevalent as these microscopic skeletons sink, that silica is typically recycled for reuse by silica-secreting organisms approximately 40 times before becoming part of the ocean-floor sediment record. The biogenic silica that becomes sediment usually crystallizes as chert.

Overall, the factors adding silica to the silica cycle include, river input, atmospheric deposition, weathering of the oceanic crust, input from hydrothermal vents, and biogenic production. Silica is removed from the cycle due to the accumulation of silica in abyssal and coastal sediments.
Silicic acid is transported to the ocean in three ways (Figure 1.1). 1) Chemical weathering of sedimentary and crystalline rocks, specifically silicate and aluminosilicate minerals, by CO₂ charged waters. This is strongly influenced by the amount of river runoff, which varies with climate. Silicic acid is then transported to the rivers either by stream or groundwater flow. Rivers also carry suspended rock fragments and particles composed of clay minerals that dissolve to create silicic acid during transport. Rivers then deposit the silicic acid into the ocean. 2) Eolian transport of particles of eroded rocks and minerals to oceanic surface waters. The particles partially dissolve upon reaching the water, resulting in the addition of silicic acid to the ocean. 3) Silicate weathering of oceanic basalt. The submarine basalt is weathered due to chemical and physical reactions from high-temperature hydrothermal activity.

Figure 1.1. Schematic of the modern silica cycle from Kidder (2000).
Large Igneous Province (LIP) activity potentially affected the Jurassic silica cycle. Separate from a few minor sea-floor volcanic pulses, three LIP intervals of mark the Latest Triassic and Jurassic (Table 1.1). The Early Jurassic (Hettangian) recovery from the Triassic-Jurassic (TJ) extinction in the aftermath of the hothouse climate that coincided with this extinction shows signs of an altered silica cycle (Ritterbush et al., 2015). These are not as severe as the apparent total shutdown of biogenic silica accumulation through the Early Triassic (e.g., Kidder and Worsley, 2004). That gap in chert formation appears to have been related to the hothouse climate that coincided with the end-Permian extinction and persisted through the Early Triassic.

One goal of this thesis is to determine if the timing of increased chert production during the Jurassic was coincident with pulses of hothouse climate triggered by LIP activity (e.g. Kidder and Worsley, 2010) as Haline Euxinic Acidic Thermal Transgression (HEATT) episodes (see Chapters 2 & 3 for further explanation). The Jurassic record provides the opportunity to relate the chert record to a spectrum of time intervals marked by significant LIP activity. The CAMP (Central Atlantic Magmatic Province) LIP marked the end of the Triassic and was probably responsible for generating that HEATT episode (Kidder and Worsley, 2010). Ritterbush et al. (2014, 2015) found that siliceous sponge-generated cherts moved from deeper marine settings into shelf environments at a time when radiolarian silica was sharply reduced just after the end-Triassic extinction.
Table 1.1. Jurassic Large Igneous Province Intervals (Large Igneous Province Commission Website)

<table>
<thead>
<tr>
<th>Large Igneous Province (ages)</th>
<th>Type</th>
<th>Location</th>
<th>Volume (v) or area (a) (millions of cubic (v) or square (a) km)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Tithonian LIPs</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Magellan Rise (150 Ma)</td>
<td>sea floor</td>
<td>Pacific Ocean</td>
<td>v = 1.8</td>
</tr>
<tr>
<td>Shatsky Rise (150 Ma)</td>
<td>sea floor</td>
<td>Pacific Ocean</td>
<td>v = 2.5</td>
</tr>
<tr>
<td>Sorachi Event (Late Tithonian)</td>
<td>sea floor</td>
<td>Pacific Ocean</td>
<td>v = 0.6</td>
</tr>
<tr>
<td><strong>Pleinsbachian-Toarcian LIP</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Karoo-Ferrar (183 Ma)</td>
<td>on land</td>
<td>South America, Africa, Antarctica</td>
<td>v = 5</td>
</tr>
<tr>
<td><strong>End-Triassic LIP</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Central Atlantic Magmatic Province (CAMP) (201 Ma)</td>
<td>on land</td>
<td>North America</td>
<td>a = 7</td>
</tr>
</tbody>
</table>

A second goal of this thesis is to establish epoch- and stage-level records of chert abundance, chert depositional environments, and the siliceous biotas that helped form these cherts. The temporal resolution of those chert records may not be short enough to recognize the typically < 1 m.y.-duration HEATT episodes, but they may provide clues to the nature of the rise of the diatoms, which seems most likely to have been during the Jurassic (Harwood and Nikolaev, 1995). Sketchy records make it unclear as to how well skeletonized the early diatoms were, but Harwood and Nikolaev (1995) have suggested they may have been weakly silicified and may have originated in nearshore settings. If the diatoms did arise in this fashion, they may have stimulated some sort of adjustment in the silica cycle. The epoch- and stage-level data assembled herein lead to hypotheses that will help refine our understanding as to whether nascent Jurassic diatoms affected the silica cycle.
CHAPTER 2: THE SIGNIFICANCE OF THE STUDY AND HYPOTHESES

2.1 Introduction

This study is a comprehensive assessment of worldwide Jurassic chert deposits. The data analyzed includes type of chert deposit (bedded versus nodular), depositional environment (peritidal versus shelf versus slope/basin), and type of siliceous skeleton included in the deposit (radiolarians versus siliceous sponges). The compilation of chert deposits is assembled by level, and Age level when possible.

The study goals were: 1) to identify and interpret Jurassic temporal and spatial patterns of marine silica accumulation in fluctuations of the marine chert abundance, 2) to report marine silica accumulation patterns with paleoclimate, paleoceanography, silica nutrient level inputs, and possible migration of siliceous organisms, and 3) to find a relationship between chert abundance and paleoenvironment conditions.

The significance of this study includes:

1. A test of a reversal in the long-term pattern of chert accumulation moving to deeper water settings suggested by Maliva et al. (1989).
2. HEATT episodes may have punctuated the long-term progression somewhat. It appears that the impact on radiolarian biotas were was sufficient to allow for brief reversions to shelf dominance of cherty facies by more primitive siliceous sponge biotas.

2.2 Hypotheses

The following hypotheses were proposed based on previous work, which will be discussed in a following chapter.
I. Large Igneous Provinces (LIP) stimulated sponge-generated shelf (or peritidal) cherts during HEATT episodes.

Based on data and hypotheses developed for the Hettangian (e.g. Ritterbush et al., 2014, 2015), Starkey and Kidder (2014) made the preliminary case that the Pliensbachian-Toarcian siliceous sponges generated conspicuous shelf cherts while the radiolarians may have been affected by a HEATT episode. Evidence for sponge-dominated shelf cherts may provide a second Jurassic example of linkage between continental LIPs and sponge-dominated shelf cherts.

II. Sea-floor LIP activity in the latest Jurassic may stimulate biogenic silica production by introducing new silica and nutrients into the ocean.

Racki and Cordey (2000) more generally suggested that deep-sea volcanic activity can potentially generate such an effect, but they did not specifically try to link such volcanism to LIP activity. The activity of three sea-floor LIPs near the end of the Jurassic (Table 1.1) provide an excellent opportunity to test this hypothesis, which is based partly on Racki and Cordey’s (2000) suggestion and the interest herein to compare effects of oceanic LIP activity to the two on-land LIPs discussed above.

III. The rise of the diatoms in the Jurassic may stimulate chert production both in terms of abundance and prevalence in nearshore settings.

It is hypothesized here that even though diatoms may have insufficient silica to preserve them well, they may still have extracted enough silica from ocean waters to generate a recognizable effect on the chert record. This hypothesis can be tested by showing that all increases in chert abundance are coincident with LIP activity.
2.3 Assumptions

The cherts in this study are Jurassic in age and are considered to be biogenically produced. Biological silica recycling produces organic chert, while inorganic silica and other nutrient sources that drive the biologic success of silica-secreting biota produce inorganic chert. Biogenic cherts are classified as deposits whose silica input is derived from the opal skeletons of siliceous organisms, and contributes toward the net output of silica in the silica cycle.

Considering that dissolution of siliceous skeletons is the dominant source of silica delivered to oceanic sediments, the abundance of chert could suggest the environments in which siliceous biotas were able to thrive. The cherts in this study represent the proximal environment of the silica-secreting organisms. Temporal variations in the chert depositional environment represent a change of the environment either because of variations in silica input or temperature.

Siliceous sponges were more important chert producers in peritidal-water settings in the Early Paleozoic (Maliva et al., 1989). New work (Kidder and Tomescu, 2016) makes the case that most of the increase in radiolarian importance in chert formation took place within the Ordovician, possibly driving the retreat of siliceous sponges and cherts from peritidal facies to shelf and deeper settings. Thus, it is assumed that radiolarians should be the dominant siliceous biota unless conditions exist to allow siliceous sponges to thrive. Although diatoms appear in the Jurassic (Sims et al., 2006), they did not influence the silica cycle significantly until the Late Cretaceous-Cenozoic (Maliva et al., 1989).
CHAPTER 3: HEATT (HALINE EUXINIC ACIDIC THERMAL TRANSGRESSION) MODEL

Icehouse, Greenhouse, and Hothouse are the three climate states that characterize Phanerozoic climate variation (e.g., Fischer, 1984; Kidder and Worsley, 2012; Hay, 2013). The Greenhouse state is considered the default state, which can be influenced to produce either the Icehouse or the Hothouse (Kidder and Worsley, 2010). During a Greenhouse state, Large Igneous Provinces (LIPs) can trigger the onset of a Haline Euxinic Acidic Thermal Transgression (HEATT) episode. A HEATT episode involves a cascade of climatic events through which a greenhouse climate shifts to a hothouse climate. HEATT episodes are expected to occur faster than the resolution provided in the record by collection of chert data. The resulting hothouse climate typically lasts less than 1 million years (Kidder and Worsley, 2012).

Bryan and Ernst (2008) define LIPs as “...magmatic provinces with areal extents >0.1 Mkm$^2$, igneous volumes >0.1 Mkm$^3$ and maximum lifespans of ~50 Myr that have intraplate tectonic settings or geochemical affinities, and are characterized by igneous pulse(s) of short duration (~1–5 Myr), during which a large proportion (>75%) of the total igneous volume has been emplaced (Fig. 3.1). The eruptions led to an epoch of environmental disruptions. LIPs are thought to emit enough SO$_4$ and CO$_2$ (up to $10^{14}$ tons of CO$_2$) to disturb the global climate and produce a significant warming effect (Kidder and Worsley, 2012). Methane can also be an important volcanix emission, depending on the nature of the crust intruded by a LIP. LIPs are linked with ocean anoxic events and
mass extinctions (Figure 3.2; Prokoph et al., 2013) both of which are characteristic of HEATT episodes.

Figure 3.1. Map of Large Igneous provinces and approximate ages of those LIPs. Modified from Coffin and Eldhom (1992).
HEATT episodes are also characterized by hot climates. Tropical sea-surface temperatures in excess of 40° C have been estimated from strata that accumulated during the early Triassic (Sun et al., 2012) and Cenomanian-Turonian (Bice et al., 2006) HEATT episodes. Sea surface temperatures at the poles can increase from 0°C during an icehouse climate to 20°C in a hothouse climate (Kidder and Worsley, 2010). The change
in temperature leads to a change in the polar-equator contrast. In an icehouse setting, the polar-equator thermal gradient is 30°C, and is estimated to be only 20°C in a hothouse climate (Kidder and Worsley, 2010). A large polar-equator contrast allows for strong winds at the scale of the planetary windbelts. Strong winds aid in global heat transport, oceanic upwelling, providing nutrients to the oceans, and increases oceanic productivity. Wind power can potentially decrease significantly during a Hothouse planetary state (Fig. 3.3) (Kidder and Worsley, 2010).

Figure 3.3. Characteristics of icehouse, greenhouse, and hothouse climates (Kidder and Worsley 2010). Climates are controlled by thermohaline circulation, cyclone distribution, latent heat distribution, sea level, wind erosive power, and ocean anoxia and euxinia.
4.1 HEATT Episodes

HEATT episodes can potentially limit biosiliceous productivity and affect the silica cycle and the chert record. For example, a chert gap has long been recognized through the Early Triassic (Erwin, 1993). This interval is probably the most extreme hothouse in the Phanerozoic. Reasons why HEATT episodes can hamper biosiliceous productivity and reduce pelagic chert accumulation include 1) changes in ocean circulation, 2) anoxia/euxinia, and 3) nutrient crisis.

A warming climate leads to weakened ocean circulation and increased ocean stratification (Kidder and Worsley, 2010). There are widespread deposits of black shale during HEATT events, which is a result of low amounts of dissolved oceanic oxygen. Ocean anoxia is among the causes of mass extinction in the oceans (Wignall, 2001). As the organisms die and settle on the ocean floor, the organic matter is incorporated into fine-grained sediment and produces deposits of black shale.

The high levels CO$_2$ emitted into the atmosphere by LIPs lead to ocean acidification (D’Hondt et al., 2004), hypercapnia (Knoll et al., 1996; Knoll et al., 2007), and calcification problems for marine organisms (Wignall et al., 2007). Because LIPs trigger HEATT episodes, there are also voluminous volcanic sulfur emissions (Adams et al., 2010), in addition to the CO$_2$ emissions, into the atmosphere and oceans during these events. The increased amount of sulfur in the anoxic oceans produces widespread euxinia, which is shown by the extensive presence of pyrite frambooids (Armstrong et al., 2009).
Anoxia and acidification affecting hothouse oceans can affect the chert record. If, for example, anoxia affects the photic zone, the productivity of silica secreting eukaryotes will be impaired. Alternatively, if anoxia affects only deeper waters, radiolarian productivity may persist in surface waters (Knoll et al., 2007). Black shale may serve to preserve radiolarians, and such shales may even be interbedded with chert. Acidification may not affect siliceous biotas, but the acid-driven shutdown of calcareous plankton could lead to more conspicuous cherts and/or black shale, if nutrient shortages do not limit biosiliceous production (Knoll et al., 2007).

Nutrient availability changes when winds are weak during a hothouse climate resulting in a reduction in upwelling occurs and productivity. Windblown dust is also reduced, which affects the transport of nutrients and oceanic life in general (Kidder and Worsley, 2010). The lack of eolian dust transport provides less iron for diazotrophs, a type of nitrogen-fixing bacteria that relies on iron (Kidder and Worsley, 2010). Negative nitrogen isotope excursions result when much of the available oceanic nitrate has been destroyed by anammox bacteria in anoxic water columns (Kidder and Worsley, 2010). If most biousable nitrogen (e.g., ammonium) is being generated by diazotrophs, a light $\delta^{15}$N excursion will result (Kidder and Worsley, 2010). This will most likely occur in warm greenhouse climates. With the onset of a hothouse, loss of iron to water-column pyrite formation will curtail the production of light $\delta^{15}$N (Kidder and Worsley, 2010). Negative nitrogen isotope anomalies (Fig. 4.1) are therefore characteristic of early onset HEATT episodes.
Potent removal mechanisms are expected to make iron scarce in hothouse oceans (Kidder and Worsley, 2010). In these chemical removal processes much of the available iron is taken up by sulfide to form the mineral pyrite. Pyrite frambooids form in the euxinic water column and then sink to the ocean floor (Wilkin and Barnes, 1997). The presence of small-diameter (< 6μm) pyrite frambooids in oceanic sediments indicates the presence of both euxinic and hothouse conditions. Such iron removal is a special case of the general tendency for many metals to be removed from ocean waters during intervals of extensive black shale deposition (Algeo, 2004).

HEATT episodes could also increase availability of ocean nutrients is to mobilize them via silicate weathering, which probably increases in a warming world (Figure 4.2) Silicate weathering delivers carbon to the oceans by converting CO$_2$ into dissolved HCO$_3^-$ as rainwater interacts with newly exposed rocks. Rivers transport nutrients and HCO$_3^-$ to

Figure 4.1. Nitrogen isotope anomalies associated with the warm greenhouse phase of HEATT events (Kidder and Worsley 2010).
the ocean, and leads to carbon burial if ocean life takes advantage of these new materials (e.g., Chamberlin, 1899; Walker et al., 1981; Moore and Worsley, 1994; Worsley et al., 1994; Berner, 2004). The burial of carbon results in global cooling. Failure to bury the carbon can lead to warming if the carbon is not sequestered in sedimentary rocks.

Silicate weathering and carbon burial can also increase when new mountain belts form (Fig. 3.2). This effect appears to be most potent during continental-collision orogenies, and large, low-latitude examples may have been important triggers for the Carboniferous and Cenozoic icehouse climates (e.g. Kidder and Worsley, 2010).

Figure 4.2. Simple schematic of the silica cycle after Kidder (2000). Arrows generalize some biogeochemical pathways for carbon, nutrients, and dissolved silica stimulated by new mountain-belt formation (e.g. continental collision orogeny).
4.2 Jurassic HEATT Episodes

There is evidence of more than 10 HEATT episodes occurring throughout the Phanerozoic (Kidder and Worsley, 2010). Each HEATT episode coincides with the formation of a LIP as well as a mass extinction event or an interval of elevated extinction intensity (Figure 4.3). The two best documented HEATT episodes during the Jurassic Period occurred at the Triassic-Jurassic boundary and the Pliensbachian-Toarcian boundary.

It is widely accepted that the Triassic-Jurassic HEATT episode was prompted by the Central Atlantic Magmatic Province (CAMP) LIP, and that the Pliensbachian-Toarcian (183-175 Ma) HEATT episode was triggered by the formation of the Karoo-Ferrar LIP (Fig. 4.3) (Pálfy and Smith, 2000; Morard et al., 2003; Wignall, 2001; Courtillot and Renne, 2003; Caruthers et al., 2013). Widespread black shale deposits throughout Europe are evidence for OAEs during these intervals (Hesselbo and Pienkowski, 2011). Additionally, high seawater temperatures are inferred from isotopic analyses (Cohen et al., 2007; Jenkyns, 2010; Jenkyns et al., 2002; McArthur et al., 2008).
Figure 4.3. Extinction intensity curves that coincide with formations of LIPs (Kidder and Worsley, 2010). Icehouse, greenhouse and hothouse climate states are also shown.
5.1 Introduction

Chert is a microcrystalline siliceous sedimentary rock that can be either organic or inorganic in origin (Hesse, 1988). Inorganic chert is formed through volcanic and hydrothermal processes, whereas organic chert is a result of biochemical or biogenic processes. “Siliceous sedimentary rocks are fine grained, dense, very hard rocks composed mainly of the of the SiO₂ minerals such as quartz, chalcedony, and opal, with minor impurities such as siliciclastic grains and diagenetic minerals” (Boggs, 1987). Chert is referred to by several different names based on qualities that are exhibited in the rock. Flint is a type of chert that occurs in a variety of colors, usually in the form of a nodule in chalk or marly limestone. Jasper is a name for chert that is red in color due to hematite inclusions. Novaculite is white to grey-black in color that is especially dense, and only found in the southern United States (Arkansas, Oklahoma, and Texas). Porcellanite is an unrefined variety of chert containing clay and calcareous matter. Silex refers to chert that is fine-grained (Tarr, 1936).

5.2 Chert Mineralogy

Amorphous silica, also known as opal-A, is the material that makes up the skeleton of siliceous organisms. Opal-A is defined by Von Rad and Rösch (1974) as highly disordered, nearly amorphous natural hydrous silica. Opal-CT is an intermediate stage of opaline silica conversion into quartz (chert). The Opal-CT occurs as microcrystalline aggregates with the crystalline structure consisting of layers of cristobalite and trydimit (Calvert, 1974). Cristobalite is a significant component of
deep-sea bedded cherts, and is thought to be an early diagenetic product of low
temperatures. Cristobalite is also referred to as $\alpha$-cristobalite, disordered cristobaline,
lussatite, opal-CT, opal-cristobalite, and unidimensionally disordered cristobalite (Wise
and Weaver, 1974).

The quartz in chert forms during the late phase of diagenesis, and is grouped by
grain size into megaquartz ($> 20 \mu m$) and microquartz ($< 20 \mu m$). Megaquartz in chert
does not contain fossils or any level of porosity (von Rad and Rösch, 1974). Microquartz
is further subcategorized by microcrystalline (1-4 $\mu m$) or chalcedonic quartz with a
fibrous habit (Carson, 1991). Chalcedonic quartz is fibrous and divided into three
categories based upon the fiber orientation. There are two length-slow types (quartzine
and lutecite), as well as one length-fast type (chalcedonite), the latter being a product of
late diagenesis (Heath, 1974).

Cherts can contain minor amounts of impurities in addition to silica, including Al,
Fe, Mn, Ca, Na, K, Mg, hematite, pyrite, or pyroclastic debris. The most abundant
constituent other than silica in chert is aluminum, followed by iron. The makeup of chert
can consist of 99% silica (pure chert) to less than 65% in certain nodular cherts (Boggs,
1987).

Two processes govern the creation of chert (Boggs, 1987). The first is
sedimentation, which includes the accumulation and concentration of biogenic opaline as
well as burial. The second is diagenesis during which the amorphous silica crystallizes.

Tréguer et al. (1995) asserted that there is a larger amount of dissolved silica
within the pore water of ocean floor sediments as opposed to overlying seawater. The
silica within the pore water remains in motion, meaning that a portion of the silica dissolves in the adjacent bottom water, some eventually solidifies as chert, and some is used as a component of aluminosilicate mineral phases. Tréguer et al. (1995) claimed that only 2.5% of all biogenically produced silica is preserved in the sediment as chert, while the rest is recycled. Heath (1974) asserted that 4% of biogenic silica is buried, but 2% is dissolved after deposition, leaving only 2% to be preserved as chert. Furthermore, the amount of dissolved silica tends to be higher within the pore waters of sediments that are rich in biogenic silica. The concentration of dissolved silica is higher also within the uppermost sediment layers, and reaches a constant concentration some tens of centimeters below the sediment surface (Calvert, 1983).

According to von Rad and Rösch (1974), the process of chert formation first begins with the remobilization of dissolved opal-A, precipitation of low temperature cristobalite, formation of microspherules (a product of early diagenesis), and finally the transformation of cristobalite into either chalcedony or micro- (> 1 µm) to cryptocrystalline (> 1 µm) quartz (a product of late diagenesis).

When silica accumulates on the ocean floor in absence of other sediment types, the result is a high opal concentration per surface unit area. The opaline skeletal material continues to dissolve after burial increasing the silica concentration in pore liquid, which results in a slow precipitation of chert. The Kastner et al. (1977) maturation theory is the most widely accepted explanation of the chert progression of diagenesis. The theory states that opal-A becomes opal-CT (cristobalite), which transforms to quartz over time. However, it is possible for opal-A to pass directly into quartz (Kastner, 1981). Calvert
(1974) asserted that the rate of the conversion of opal-CT to quartz depends exclusively on temperature.

The rate of chert diagenesis of chert during lithification is controlled by several factors. These factors include in situ temperature (Li, 1995; Carson, 1991; Calvert, 1974; Kastner et al., 1977), chemistry of the pore fluid, pH, time, mineralogy of the host rock, abundance of organic matter, clay minerals, burial depth, porosity and permeability of the host rock or sediment, and specific surface area (Williams et al., 1985; Tribble et al., 1995). According to von Rad and Rösch (1974), “under normal deep-sea conditions, time appears to be the single most important parameter” (p. 344). In an ideal setting for diagenesis, a sedimentary column would contain opal-A overlain by opal-CT, which would overlie quartz. Between each constituent would be a transition zone of mixed phases (Williams et al., 1985). In an experiment conducted by Kastner et al. (1977), the factors that most significantly affect the conversion of opal-A to opal-CT were time, temperature, host rock, and solution composition. Their study showed that the rate of the conversion increased when the host rock is carbonate, as opposed to host sediments that are rich in clay. Alkaline and magnesium-rich solutions also increased the conversion rate. Magnesium that comes from the seawater and hydroxide provided by calcite dissolution become hydroxide sites for the nucleation of opal-CT, and later the growth of opal-CT lepispheres. The connection between carbonate and chert formation is due to alkalinity. The pore water solutions dissolve the calcite and become alkaline, which enhances the conversion of opal-A into opal-CT at a faster rate. Kastner et al. (1977) suggested that clay minerals, like smectites, compete with the opal-CT for the alkalinity
in the seawater, and therefore slow the opal-A to opal-CT transformation and make the clay minerals magnesium-rich.

5.3 Chert Morphology

Chert exists in two forms: bedded and nodular. Bedded and nodular cherts differ either morphologically (Boggs, 1987) or genetically. Bedded cherts are regarded to be the result of the accumulation of planktonic silica-secreting biota, such as radiolarians (radiolarites) and diatoms (diatomites) (Grunau, 1965; Aubouin, 1965; Ramsay, 1973; Hein and Parrish, 1987; Maliva and Siever, 1989) and are typically found in siliceous host rocks (Wise and Weaver, 1974). Nodular cherts are considered to be the product of diagenesis resulting from the remobilization of silica and mineral precipitation within a limestone host rock (Maliva and Siever, 1989; Maliva et al., 1989).

Bedded cherts can form in both deep water and peritidal environments but is directly influenced by planktonic activity, which is influenced by oceanic conditions (circulation, nutrient supply, etc.), sediment input, and depth. Modern silica is most abundant below the CCD in deep-water settings, where siliceous oozes form that are made exclusively of biogenic silica from radiolarians and diatoms. Bedded cherts can also form in peritidal environments, provided the surface water is rich in nutrients, there is a lack of calcareous organisms, and there is little terrestrial sedimentation.

Whether bedded chert is produced inorganically or biogenically is a topic of debate among geologists. Some scientists argue for an inorganic origin, especially for cherts within a volcanic sequence. The presence of minerals that are known to be
products of volcanism (montmorillonite, palygorskite, sepiolite, and cinoptilolite) within bedded chert support the theory of an inorganic origin (Wise and Weaver, 1974).

The “gel theory” suggests that chert is formed by a siliceous gel that collects radiolarian remains during the solidification of the gel (Davis, 1918). The cause of high accumulations of siliceous tests produced by silica-secreted planktonic biota is a product of debate. Possible mechanisms include: blooms of plankton induced by submarine volcanism (Aubouin, 1965), an accumulation of siliceous material below the CCD that is not caused by volcanism (Aubouin, 1965; Garrison, 1974; Heath, 1974), and high productivity of silica-secretting planktonic organisms induced by upwelling zones (Calvert, 1966; Ramsay, 1973; Hein and Parrish, 1987).

Bramlette (1946) claimed that siliceous oozes, created due to biogenic silica accumulation, mature from porcelanite to chert. Bramlette’s theory is supported by the alteration of opal-A, which is considered to be a depth-, temperature-, and time-dependent process (Wise and Weaver, 1974; von Rad and Rösch, 1974).

Chert nodules can greatly vary in size and shape, from spherical to irregular shapes. It is even possible for the nodules to occasionally conjoin to form layers similar bedded chert. The nodules may or may not be laminated or fossiliferous. Nodular chert forms primarily in carbonate host lithologies, and in both peritidal and deep water depositional settings (Maliva and Siever, 1989). “There is no intrinsic depositional environment restriction on nodular chert formation,” (Maliva and Siever, 1989, p.425), including hypersaline, marine, and mixed marine-meteoric pore waters. Chert nodules have also been known to form in mudrocks, evaporites, burrows, and can even nucleate
around fossils, although rarely. For the most part, nodules form contemporaneously with the bedding planes. Kastner et al. (1977) were able to explain experimentally the relationship between carbonate and the precipitation of silica.

According to Wise and Weaver (1974), there are two theories as to how nodular chert forms. The first is the maturation theory, and the second is the quartz precipitation theory. The maturation theory suggests that a chert nodule begins as disordered cristobalite that transforms into chalcedonic quartz over time starting from the center of the nodule (Heath and Moberly, 1974. Lancelot (1973a) introduced the quartz precipitation theory, which suggests that a chert nodules begins with the precipitation of quartz and that disordered cristobalite is a by-product. This theory is valid only provided that the host lithology is a carbonate ooze sequence and the mineralogy of the chert is comprised solely of quartz. The theory is broken down into three steps. First, the quartz precipitates; second, growth of the quartz assuming highly permeable sediment; and finally, the crystallization of the cristobalite at the outer edges of the nodule.

Wise and Weaver (1974) tested these two theories using scanning electron microscopy. They were able to conclude that there are seven steps to the formation of nodular chert: 1) diffusion of fluid supersaturated with silica; 2) precipitation of cristobalite; 3) dissolution of carbonate host rock with contemporaneous growth of a nodule nucleus; 4) growth of nodule; 5) transformation of cristobalite to quartz starting at the nucleus and continuing outward; 6) pore space filled with chalcedony and cryptocrystalline quartz; and 7) continued growth until dissolved silica is depleted.
Maliva and Siever (1989) found that none of the previous models of nodular chert evolution, were consistent with their observations. The authors maintain that the rate of siliceous precipitation must be equal to the rate of carbonate dissolution. They call this model the force of crystallization-controlled replacement model. An undersaturation of calcite is restricted to the contact of silica-calcite, meaning calcite only dissolves where silica is present. Their explanation is valid in all depositional environments, and gives rise to the idea that the only constituent required to form a nodular chert is pore water supersaturated with quartz or opal-CT. According to Maliva and Siever (1989), there are three attributes of the host sediment that potentially affect the formation of nodular chert, as well as the distribution of the nodules within the bedding planes. These three attributes are amount of organic matter, porosity and permeability, and concentration of biogenic opal.

Von Rad and Rösch (1974) claimed that biogenic siliceous sediment takes 70-90 My to transform into a mature quartz chert. Siliceous oozes can become cherts in 25-50 My, according to Kastner et al. (1977). However, the time for opal-A to convert to quartz will decrease in the formation temperature increases. Quartz can form within 30 My provided the *in situ* temperature is approximately 30°C (Heath, 1973). These age estimates only apply when the siliceous solution completes the entire maturation process, not if quartz precipitates directly from solution.

5.4 Chert Abundance in the Geologic Record

Patterns in chert abundance have been examined and compiled in previous studies (Grunau, 1965; Dietz and Holden, 1966; Ramsay, 1973; Ronov, 1982; Hein and Parrish,
Grunau (1965) suggested that chert abundance could be related to the formation of ophiolites. Grunau found a correlation between thicknesses of coeval radiolarian chert and ophiolite deposits and suggested that the silica was produced volcanically due to the eugeosynclinal position of several of the radiolarites. Work by Racki and Cordey (2000) indicated that volcanically derived silica sources were more prevalent in the Paleozoic. It is possible that the volcanic activity supplied not only silica, but also other nutrients. That would suggest that volcanic cherts might not be simply inorganic. The nutrients provided to the oceans by the volcanic activity could have fed siliceous biota, and therefore stimulated chert production.

Ramsay (1973) and Ronov (1982) made compilations of chert abundance through time grouped by period that were used by Hein and Parrish (1987) to create a paleo-upwelling model to explain of bedded chert distribution. Because radiolarians are sensitive to environmental factors such as water temperature and chemistry, they can serve as a paleoceanographic proxy. The number and environment of modern radiolarians is distinctly related to the nutrients available, as well as depth (Kruglikova, 1993; Abelmann and Gowig, 1997). Hein and Parrish (1987) indicated that radiolarian abundance in chert deposits can be used to suggest deep ocean environments, but can also be correlated with moderate depth environments (Aubouin, 1965; Ormiston, 1993). Changes in radiolarian diversity can be attributed to variations in sea level, paleoceanographic regime, and plate tectonics (Vishnevskaya, 1997).

Kidder and Erwin (2001) examined the distribution of Phanerozoic bedded chert to determine if siliceous facies were affected by climate changes and the resulting
extinctions. They further developed the work by Hein and Parrish (1987) by separating data by, as well as differentiating basinal versus shelf cherts by facies associations. The chert abundance data shows that bedded cherts occurrences almost double in the Jurassic compared to the Triassic. Furthermore, when Kidder and Erwin (2001) normalized the raw chert data to outcrop area/time (after Raup, 1976), the is a four-fold increase in bedded cherts in the Jurassic. The Jurassic chert peaks might be associated with the abundance of LIPs during the Jurassic that would have increased the amount of silicate weathering, in turn increasing the terrestrial silica input into the oceans. Also, the overall abundance of chert in the Jurassic might be partly a function of the extensive ocean spreading associated with sea-floor volcanism after the rifting of Pangea during the Jurassic. Although Cretaceous cherts were numerically more abundant than Jurassic cherts, Jurassic cherts are more abundant than Cretaceous cherts when the Jurassic cherts are normalized to outcrop area/time.

Maliva et al. (1989), Kidder and Erwin (2001), and Kidder and Mumma (2003) examined the prevailing environment of Phanerozoic siliceous organisms, as well as their migration patterns. Maliva et al. (1989) proposed that nodular and bedded chert formation from the Silurian to the Early Cretaceous was dominant in deep ocean environments, and common in subtidal shelf and platform environments, but uncommon in peritidal marine settings.
CHAPTER 6: METHODS

Compilation of worldwide Jurassic chert deposits (bedded and nodular) generated a high resolution of results at both the series/epoch level and at the stage/age level. The compilation of deposits is based on several sources, including GeoRef, Science Citation Index Expanded, and geology journals. Despite temporal resolution that is usually as fine as stage-level in this study, the HEATT episodes that are part of the focus of this work are not recognized in systematic chert tabulations. This is partly because they are more rapid than the duration of the stages. It may also be partly because the HEATT effects on the silica cycle do not, as yet, appear in the global picture. HEATT effects do appear pertinent in some locations in which the facies are of a nature to clearly show the HEATT effects. This issue is addressed in more depth in the Discussion section.

Chert data were managed using Microsoft Office Excel for Mac 2011. Where available the information collected included: 1) the geographic location (country, region), including coordinates (latitude and longitude); 2) the geologic age at the epoch (Lower/Early, Middle, and Upper/Late Jurassic) or epoch level (e.g. Toarcian, Tithonian); 3) the name of the chert formation; 4) associated lithologies; 5) the color of the chert deposit; 6) the type of chert deposit (bedded or nodular); 7) the depositional environment (peritidal, shelf, or slope/basin); 8) the type of fossils included in the deposit (radiolarian or siliceous sponge); 9) general comments and notes; and 10) the bibliographic reference.

The data collected in order to determine: 1) age of chert at level (Lower, Middle, or Upper Jurassic), and at the stage/age level (e.g. Toarcian, Tithonian) when data were
available; 2) type of chert deposition (bedded or nodular); 3) depositional environment of chert deposits (peritidal, shelf, or slope/basin); 4) deduce the determining factors of chert genesis.

Chert occurrences were also collected from detailed sedimentologic and stratigraphic papers, which often contained reliable data relating to the chert deposits, such as age, origin, fossil content, and depositional environment. Other documented chert occurrences were collected from other geological studies that mentioned Jurassic chert, but the study did not focus on the chert specifically. In these cases, certain factors were inferred from the data available. For example, in the absence of other information, bedded radiolarian cherts interbedded with black shale was interpreted as a slope/basin setting. Still, determining depositional environment was not always possible from the information available on a given chert deposit and some depositional environments appear on data figures as undetermined.

The types of chert deposits considered in this study were nodular cherts, bedded cherts, and type undetermined. The difference between bedded and nodular cherts was considered to be only a physical difference, as opposed to a difference in style of formation.

Most geologic studies used to create the Jurassic chert compilation included depositional environment, which was determined by the author based on paleogeography, lithologic associations, sedimentary structures, and other environmental reconstruction criteria. Depositional environments were grouped into four categories: 1) peritidal; 2) shelf; 3) slope/basin; 4) undetermined.
The paleogeography of the in situ chert deposits is especially helpful with respect to determining the depositional setting. Both clastic and carbonate platforms suggest peritidal to shelf environments in a miogeoclinal settings. When the chert deposits are associated with volcanic lithologies, the paleogeography is important in helping to determine if the depositional environment was deep water (away from the craton) or closer to peritidal water settings (near the craton). Depositional environment indicators, such as sedimentary structures and lithological associations, are necessary tools when the depositional environment is unknown.

It should be considered that there are certain biases when it comes to determining the depositional setting of chert deposits. Cherts associated with limestone lithologies are interpreted having a peritidal or shelf depositional setting. This is probably a safe assumption in the Jurassic. Although calcareous plankton were appearing at about this time, deep-water, plankton-generated carbonates do not appear to be of major significance in the geologic record until the Cretaceous, when the chalks become a significant component of the pelagic record. Contrastingly, cherts associated with shale lithologies are interpreted as having a deep-water depositional environment. Although these assumptions predominantly hold true, it is unavoidable that there are some exceptions.
CHAPTER 7: RESULTS

The compilation of worldwide chert deposits yielded 87 occurrences spanning 5 continents (North America, South America, Europe, Asia, and Australia) and 37 countries. Each occurrence is listed in the appendix. Some cherts span epoch boundaries (e.g. Early and Middle Jurassic) resulting in some double counting for epoch-level analysis. This inflates the 87 cherts in the appendix to an apparent total of 107 cherts. The effect of these minor levels of double-counting does not appear significant enough to affect the interpretations based on chert abundance in each epoch.

7.1 Jurassic Chert Abundance by Epoch

Table 7.1 presents chert counts differentiated by epoch, as well as a normalized frequency relative to the duration of the epoch. The Late Jurassic is characterized by the most chert relative to the Early and Middle epoch. The highest chert count occurs in the Late Jurassic. It is also important to note that the Late Jurassic was the shortest epoch in the Jurassic, making the high amount of chert even more significant.

The Middle Jurassic exhibits the second-highest number of cherts, whereas the data suggest that the Early Jurassic contains the least amount of chert relative to the rest of the Jurassic Period (Figure 7.1).
Table 7.1. Chert by Epoch Level

<table>
<thead>
<tr>
<th><strong>Epoch</strong></th>
<th><strong>Count</strong></th>
<th><strong>Duration (Ma)</strong></th>
<th><strong>N/Ma</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td>Early</td>
<td>22</td>
<td>24</td>
<td>0.9</td>
</tr>
<tr>
<td>Middle</td>
<td>32</td>
<td>18</td>
<td>1.78</td>
</tr>
<tr>
<td>Late</td>
<td>45</td>
<td>12</td>
<td>3.75</td>
</tr>
<tr>
<td>Undetermined</td>
<td>8</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>107</strong></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 7.1. Pie chart of frequency and relative percentage of chert by epoch level. Whole integers note the raw numerical abundance of chert per epoch from Table 10.1. Percentages are based on those raw abundance values normalized to the total number of cherts (107).
7.2 Jurassic Chert Abundance by Stage

Figure 10.2 is a bar graph of chert frequency by stage. The Callovian and Bathonian stages are the most chert-abundant with 22 and 21 occurrences, respectively. The Aalenian stage exhibits the least amount of chert, with only 6 documented occurrences. It is important to note that there is a general increasing trend of chert frequency during the Jurassic.

CHERT BY STAGE LEVEL

<table>
<thead>
<tr>
<th>Stage</th>
<th>Frequency</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tithonian</td>
<td></td>
</tr>
<tr>
<td>Kimmeridgian</td>
<td></td>
</tr>
<tr>
<td>Oxfordian</td>
<td></td>
</tr>
<tr>
<td>Callovian</td>
<td></td>
</tr>
<tr>
<td>Bathonian</td>
<td></td>
</tr>
<tr>
<td>Bajocian</td>
<td></td>
</tr>
<tr>
<td>Aalenian</td>
<td></td>
</tr>
<tr>
<td>Toarcian</td>
<td></td>
</tr>
<tr>
<td>Pleinsbachian</td>
<td></td>
</tr>
<tr>
<td>Sinemurian</td>
<td></td>
</tr>
<tr>
<td>Hettangian</td>
<td></td>
</tr>
</tbody>
</table>

Figure 7.2. Number of chert occurrences by stage level.
7.3 Type of Jurassic Chert Deposition at the Epoch Level

A breakdown of chert depositional type (bedded, bedded and nodular, or nodular) is displayed in Table 7.3. Each epoch is comprised of mostly bedded chert with few instances of both bedded and nodular chert. Only the Late Jurassic is characterized with nodular cherts unassociated with a bedded chert. The highest relative percentage of bedded chert exists in the Middle Jurassic, while the highest percentage of both bedded and nodular chert exists in the Early Jurassic. The author found 3 instances of bedded chert and was unable to determine the associated epoch.

<table>
<thead>
<tr>
<th>Epoch</th>
<th>Bedded</th>
<th>%</th>
<th>N/Ma</th>
<th>Bedded &amp; Nodular</th>
<th>%</th>
<th>N/Ma</th>
<th>Nodular</th>
<th>%</th>
<th>N/Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td>Early</td>
<td>14</td>
<td>87.5</td>
<td>0.58</td>
<td>2</td>
<td>12.5</td>
<td>0.08</td>
<td>0</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>Middle</td>
<td>27</td>
<td>93</td>
<td>1.5</td>
<td>2</td>
<td>7</td>
<td>0.1</td>
<td>0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Late</td>
<td>26</td>
<td>81</td>
<td>2.17</td>
<td>1</td>
<td>3</td>
<td>0.08</td>
<td>5</td>
<td>1</td>
<td>0.417</td>
</tr>
<tr>
<td>Undetermined</td>
<td>3</td>
<td>100</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>5</td>
<td>5</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

7.4 Type of Jurassic Chert Deposition at the Stage Level

Figure 7.4 shows chert depositional type (bedded, bedded and nodular, or nodular) at the stage level. The graph clearly shows that bedded cherts are the most common depositional type Jurassic-wide. Occurrences of both bedded and nodular cherts are much less common, but do occur at least once in every stage of the Jurassic. Nodular cherts do not occur independently of bedded cherts until the Late Jurassic.
Figure 7.4. Number of chert occurrences during the Jurassic differentiated by stage and chert type.

7.5 *Jurassic Chert Depositional Environment Examined at the Epoch Level*

Figure 7.5 shows frequency of depositional environment (deep, shelf, or peritidal) per epoch. The graph shows that each epoch preserves a significant amount of chert occurrences with an undeterminable depositional environment. Both the Early and Middle Jurassic are most abundant in shelf-generated cherts, with the Middle Jurassic exhibiting the most shelf cherts overall. The Late Jurassic is most abundant in deep-water cherts.
Figure 7.5. Number of chert occurrences differentiated by epoch and depositional environment.

7.6 Jurassic Chert Depositional Environment Examined at the Stage Level

The following chart depicting depositional environment by stage (Fig. 7.6) shows that shelf cherts dominated the depositional environment (as opposed to peritidal or deep) until the Oxfordian stage in the Late Jurassic. The Oxfordian through the Tithonian were dominated by deep-water cherts, although shelf cherts were still relatively abundant. Although the number of cherts with an undetermined depositional environment could render the frequency difference between shelf and deep-water settings to subequal
proportions for many of the epoch, it is safe to say that these two settings are the dominant ones for Jurassic cherts as Maliva et al. (1989) predicted. Peritidal cherts have been documented at every Jurassic stage expect during the Aalenian, but were proportionally more abundant during the Early Jurassic. Such shallow-water cherts were unexpected in Jurassic strata.

DEPOSITIONAL ENVIRONMENT BY STAGE

![Graph showing frequency of chert occurrences by stage and depositional environment]

FREQUENCY

Figure 7.6. Number of chert occurrences differentiated by depositional environment and stage.

7.7 Jurassic Chert Fossil Type at the Epoch Level

Radiolarian-abundant cherts were the most common chert type throughout all of the Jurassic (Fig. 7.7). Siliceous sponge cherts were the most frequent during the Middle
Jurassic, and still relatively abundant during the Early Jurassic. Siliceous sponge cherts were the least relevant during the Late Jurassic, when radiolarian cherts dominated. Although diatoms appeared in the Jurassic record (e.g. Sims et al., 2006; Harwood and Nikolaev, 1995), none of the literature used in this study reported the presence of diatoms in chert.

Figure 7.7. Occurrences of chert sorted by epoch and fossil type.
7.8 *Jurassic Chert Fossil Type at the Stage Level*

Every stage is dominant in radiolarian cherts, while siliceous sponge cherts increase toward the Bajocian, where they remained at their highest abundance during the Jurassic until after the Callovian. By the Oxfordian, siliceous sponge cherts began to decrease, and saw a slight increase during the Tithonian.

**Fossil Type by Stage**

![Fossil Type by Stage](image)

**Figure 7.8.** Occurrences of chert sorted by stage and fossil type.

7.9 *Fossil Type by Depositional Environment Organized by Epoch*

Table 7.9 shows that Early and Middle Jurassic cherts were dominantly radiolarian cherts deposited in shelf environments. Conversely, the Late Jurassic cherts were radiolarites mostly deposited in deep marine settings. Siliceous sponges tended to
be most common in shelf and peritidal settings, with nine of the eleven documented sponge-bearing cherts being either shelf (8 cherts) or peritidal (1 chert). Siliceous sponges were reported in only two deep-water cherts.

Table 7.9. Fossil Type and Depositional Environment by Epoch

<table>
<thead>
<tr>
<th>Epoch</th>
<th>Fossil Type</th>
<th>Peritidal</th>
<th>Shelf</th>
<th>Deep</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Late</td>
<td>R</td>
<td>1</td>
<td>3</td>
<td>8</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>SS</td>
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<td>1</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>U</td>
<td>3</td>
<td>3</td>
<td>2</td>
<td>8</td>
</tr>
<tr>
<td></td>
<td>Total</td>
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<td>7</td>
<td>10</td>
<td></td>
</tr>
<tr>
<td>Middle</td>
<td>R</td>
<td>1</td>
<td>5</td>
<td>6</td>
<td>12</td>
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<td>7</td>
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<td>U</td>
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<td>6</td>
</tr>
<tr>
<td></td>
<td>Total</td>
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<td>14</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>Early</td>
<td>R</td>
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<td>6</td>
<td>1</td>
<td>10</td>
</tr>
<tr>
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<td>SS</td>
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<td>1</td>
<td>3</td>
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<td></td>
<td>U</td>
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<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td>3</td>
<td>8</td>
<td>2</td>
<td></td>
</tr>
</tbody>
</table>

7.10 Associated Facies of Early Jurassic Chert and Average Latitude

Table 7.10 demonstrates that cherts of the Early Jurassic were most commonly associated with limestone, shale, and sandstone, respectively. The majority of the Early Jurassic chert deposits were deposited between present-day latitudes of 23°N and 34°N. Mudstone and volcanic facies were less abundant but still noteworthy. Less significant depositional facies include ophiolites, breccias, conglomerates, and siltstones.
Table 7.10. Early Jurassic Associated Facies

<table>
<thead>
<tr>
<th>Associated Facies</th>
<th>Count</th>
<th>%</th>
<th>Average Latitude (°N)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Limestone</td>
<td>12</td>
<td>32</td>
<td>34</td>
</tr>
<tr>
<td>Shale</td>
<td>7</td>
<td>18</td>
<td>23</td>
</tr>
<tr>
<td>Sandstone</td>
<td>6</td>
<td>16</td>
<td>28</td>
</tr>
<tr>
<td>Volcanic</td>
<td>4</td>
<td>10.5</td>
<td>50</td>
</tr>
<tr>
<td>Mudstone</td>
<td>4</td>
<td>10.5</td>
<td>23</td>
</tr>
<tr>
<td>Ophiolite</td>
<td>2</td>
<td>5</td>
<td>38</td>
</tr>
<tr>
<td>Breccia</td>
<td>1</td>
<td>3</td>
<td>41</td>
</tr>
<tr>
<td>Conglomerate</td>
<td>1</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>Siltstone</td>
<td>1</td>
<td>2</td>
<td>10</td>
</tr>
<tr>
<td>Marble</td>
<td>0</td>
<td>0</td>
<td>n/a</td>
</tr>
</tbody>
</table>

7.11 Associated Facies of Middle Jurassic Chert and Average Latitude

Middle Jurassic associated facies of chert deposits were mostly limestone, volcanics, and shale. These deposits are currently located at latitudes between 35 and 43 °N on average. Cherts deposited with mudstones are also somewhat significant, and exist at an average latitude of 37°N. Least significant facies were breccias, sandstones, and ophiolites.
Table 7.11. Middle Jurassic Associated Facies

<table>
<thead>
<tr>
<th>Associated Facies</th>
<th>Count</th>
<th>%</th>
<th>Average Latitude (°N)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Limestone</td>
<td>13</td>
<td>32.5</td>
<td>40</td>
</tr>
<tr>
<td>Volcanic</td>
<td>11</td>
<td>27.5</td>
<td>35</td>
</tr>
<tr>
<td>Shale</td>
<td>6</td>
<td>15</td>
<td>43</td>
</tr>
<tr>
<td>Mudstone</td>
<td>5</td>
<td>12.5</td>
<td>37</td>
</tr>
<tr>
<td>Breccia</td>
<td>2</td>
<td>5</td>
<td>41</td>
</tr>
<tr>
<td>Sandstone</td>
<td>2</td>
<td>5</td>
<td>44</td>
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<tr>
<td>Ophiolite</td>
<td>1</td>
<td>2.5</td>
<td>41</td>
</tr>
<tr>
<td>Conglomerate</td>
<td>0</td>
<td>n/a</td>
<td>n/a</td>
</tr>
<tr>
<td>Siltstone</td>
<td>0</td>
<td>n/a</td>
<td>n/a</td>
</tr>
<tr>
<td>Marble</td>
<td>0</td>
<td>n/a</td>
<td>n/a</td>
</tr>
</tbody>
</table>

7.12 Associated Facies of Late Jurassic Chert and Average Latitude

The most abundant cherts associated facies during the Late Jurassic were limestones and volcanics, with latitudes between 34 and 35 °N on average. Still noteworthy, sandstones and shales associated with Late Jurassic cherts are at average latitudes between 34 and 37°N. Least significant associated facies are breccias, siltstones, mudstones, ophiolites, conglomerates and marble.
Table 7.12. Late Jurassic Associated Facies

<table>
<thead>
<tr>
<th>Associated Facies</th>
<th>Count</th>
<th>%</th>
<th>Average Latitude (°N)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Limestone</td>
<td>21</td>
<td>30</td>
<td>35</td>
</tr>
<tr>
<td>Volcanic</td>
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<tr>
<td>Shale</td>
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<td>11</td>
<td>37</td>
</tr>
<tr>
<td>Sandstone</td>
<td>7</td>
<td>10</td>
<td>34</td>
</tr>
<tr>
<td>Breccia</td>
<td>4</td>
<td>6</td>
<td>45</td>
</tr>
<tr>
<td>Siltstone</td>
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<td>Mudstone</td>
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<tr>
<td>Ophiolite</td>
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<td>50</td>
</tr>
<tr>
<td>Conglomerate</td>
<td>1</td>
<td>1.5</td>
<td>60</td>
</tr>
<tr>
<td>Marble</td>
<td>1</td>
<td>1.5</td>
<td>46</td>
</tr>
</tbody>
</table>

7.13 Colors and Depositional Environment of Early Jurassic Cherts

The most common chert colors during the Early Jurassic were red and gray. Both of these colors occurred most frequently in shelf depositional environments, but also occurred in peritidal and deep environments. The average latitude for the red cherts is 35°N and 37°N for the gray cherts.

The green cherts are also abundant during the Early Jurassic, and mostly deposited in shelf environments. One green chert has been documented as being deposited in a peritidal environment, and there are none known to have been deposited in a deep environment. The average latitude for Early Jurassic green cherts is 38°N.

Black cherts were also somewhat significant during the Early Jurassic. The three occurrences were equally distributed in peritidal, shelf, and deep environments. The average latitude for these cherts is 35°N.
Table 7.13. Early Jurassic Chert Colors

<table>
<thead>
<tr>
<th>Color</th>
<th>Count</th>
<th>%</th>
<th>Average Latitude (°N)</th>
<th>Peritidal</th>
<th>Shelf</th>
<th>Deep</th>
</tr>
</thead>
<tbody>
<tr>
<td>Red</td>
<td>7</td>
<td>28</td>
<td>35</td>
<td>1</td>
<td>4</td>
<td>1</td>
</tr>
<tr>
<td>Gray</td>
<td>7</td>
<td>28</td>
<td>37</td>
<td>1</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>Green</td>
<td>4</td>
<td>16</td>
<td>38</td>
<td>1</td>
<td>3</td>
<td>0</td>
</tr>
<tr>
<td>Black</td>
<td>3</td>
<td>12</td>
<td>35</td>
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<td>Brown</td>
<td>1</td>
<td>4</td>
<td>37</td>
<td>0</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>White</td>
<td>1</td>
<td>4</td>
<td>35</td>
<td>1</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Yellow</td>
<td>1</td>
<td>4</td>
<td>34</td>
<td>0</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>Pink</td>
<td>1</td>
<td>4</td>
<td>46</td>
<td>0</td>
<td>1</td>
<td>0</td>
</tr>
</tbody>
</table>

7.14 Colors and Depositional Environment of Middle Jurassic Cherts

Middle Jurassic cherts are similar to the Early Jurassic occurrences in that the most common chert colors are red and gray. The average latitude of these colors is 37°N and 40°N, respectively. The red cherts were mostly deposited in shelf environments, closely followed by deep environment occurrences. There is one known occurrence of a red chert with a peritidal depositional environment during the Middle Jurassic. The gray cherts were dominated by shelf depositional environments, closely followed by peritidal depositional environments. There is one deep environment occurrence of gray chert during the Middle Jurassic.

Green and brown cherts were equally significant during the Middle Jurassic, both exhibiting deposition in both peritidal and shelf environments. The average latitude of the green cherts is 34°N, while the average latitude of the brown cherts is 29°N.
Table 7.14. Middle Jurassic Chert Colors

<table>
<thead>
<tr>
<th>Color</th>
<th>Count</th>
<th>%</th>
<th>Average Latitude (°N)</th>
<th>Peritidal</th>
<th>Shelf</th>
<th>Deep</th>
</tr>
</thead>
<tbody>
<tr>
<td>Red</td>
<td>14</td>
<td>45</td>
<td>37</td>
<td>1</td>
<td>5</td>
<td>4</td>
</tr>
<tr>
<td>Gray</td>
<td>7</td>
<td>23</td>
<td>40</td>
<td>2</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>Green</td>
<td>3</td>
<td>10</td>
<td>34</td>
<td>0</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Brown</td>
<td>3</td>
<td>10</td>
<td>29</td>
<td>0</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Pink</td>
<td>2</td>
<td>6</td>
<td>47</td>
<td>0</td>
<td>2</td>
<td>0</td>
</tr>
<tr>
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<td>1</td>
<td>0</td>
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<td>0</td>
</tr>
</tbody>
</table>

7.15 Colors and Depositional Environment of Late Jurassic Cherts

The dominant chert colors during the Late Jurassic were red and green. Both of these colors were mostly deposited in deep depositional environments. Both the red and green cherts also have an average latitude of 38°N.

Gray cherts during the Late Jurassic are also significant, mostly deposited in deep setting, with one occurrence each in peritidal and shelf settings. The average latitude of gray cherts is 42°N.
Table 7.15. Late Jurassic Chert Colors

<table>
<thead>
<tr>
<th>Color</th>
<th>Count</th>
<th>%</th>
<th>Average Latitude (°N)</th>
<th>Peritidal</th>
<th>Shelf</th>
<th>Deep</th>
</tr>
</thead>
<tbody>
<tr>
<td>Red</td>
<td>10</td>
<td>31</td>
<td>38</td>
<td>0</td>
<td>2</td>
<td>5</td>
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<td>3</td>
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<tr>
<td>Gray</td>
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<td>Brown</td>
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<td>12.5</td>
<td>36</td>
<td>1</td>
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<td>3</td>
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<td>n/a</td>
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<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Pink</td>
<td>0</td>
<td>0</td>
<td>n/a</td>
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</tr>
</tbody>
</table>
CHAPTER 8: RADIOLARIANS AND SILICEOUS SPONGES DURING THE JURASSIC

8.1 Radiolarians

Radiolarians are classified as marine single-celled planktonic organisms with siliceous skeletons. Only polycystine radiolarians have opaline silica skeletons and preserve well as fossils. Polycystine radiolarians are divided into spumellar and nasellar radiolarians. Spumellar radiolarians have a spherical symmetry, whereas nasellar radiolarians have a bilateral symmetry.

Radiolarians are among the microorganisms that are capable of thriving in a wide spectrum of environments, from deep basin environments to peritidal water environments. Radiolarians exist in various shapes, and range in size from 30 microns to 2 millimeters in diameter. They acquire nutrients by acting as predators, filter feeding, or living in symbiotic relationships with unicellular algae like dinoflagellates (Casey, 1993). The majority of radiolarians live as solitary forms, but it has been documented that some species tend to form colonies (Casey, 1993).

Currently, most radiolarians inhabit warm surface waters that are nutrient rich, usually within the equatorial zone. Radiolarians also less commonly exist in tropical and subtropical settings. They choose their environments based on salinity, temperature, nutrient availability, as well as other parameters. Due to their sensitivity to these parameters, radiolarians serve as a paleoceanographic proxy to paleoclimate conditions. The amount and location of modern day radiolarians are highly associated with oceanic areas of nutrient-rich conditions, as well as depth (Abelmann and Gowing, 1997;
Kruglikova, 1993; Chang et al., 2003). Average oceanic temperature that modern radiolarians thrive is 27°C, with a salinity of 35-40‰ (Anderson et al, 1989a, b, c; 1990; Matsuoko and Anderson, 1992; Sugiyama and Anderson, 1997). Hein and Parrish (1987) recognized that frequency of radiolarian-bearing formations serve as a proxy for paleoupwelling and areas of high productivity and high nutrients. It has also been documented that radiolarians evolution was a response to changes in sea level and plate tectonic events (Vishnevskaya, 1997).

8.2 Siliceous Sponges

Sponges are grouped into four classes, largely by sponge skeleton morphology. The four classes consist of Demospongea, Hexactinellidea, Calcarea, and Sclerospongea (Rigby, 1983). Classes Hexactinellidae and Demospongea are comprised of the siliceous spicules seen in fossilized siliceous sponges. Sponges are immobile metazoans that are filter-feeding organisms. Demosponges currently comprise of about 95% of all types of sponges (Rigby, 1983). There are more the 600 genera of Demospongea that are capable of living in any oceanic setting (Prothero, 2004). The majority of modern sponges inhabit peritidal marine environments. However, in general, modern siliceous sponges favor marine settings as deep as 5000 meters (Prothero, 2004). It is important to note that there are some Demospongea that currently inhabit peritidal water settings in latitudes as high as 55° N, and cold temperatures (10°C) (Reincke and Barthel, 1997). Hexactinelliea tend to live at depths ranging from 200 to 2000 meters, but also exist in deep abyssal environments 6000 meters and deeper (Noble, 1997; Tabachnik, 1994; and Hartman, 1983).
Demosponges and Hexactinellids each require a different amount of dissolved silica for survival, as evident by where these sponges are respectively found most commonly. For example, Hexactinellids require higher amount of Si(OH)$_4$ as opposed to Demosponges. Present day Hexactinellids live in deep marine settings that exhibit a high amount of dissolved silica (Tréguer et al., 1995; Nelson et al., 1995), as well as peritidal habitats with silica-rich waters. It has even been documented that Hexactinellids exist in microhabitats that receive silica from freshwater sources (i.e., rivers, streams, etc.) (Maldonado et al., 1999; Boury-Esnault and Vacelet, 1994).

Maldonado et al. (1999) explain that siliceous sponges require higher levels of dissolved silica in order to thrive as opposed to radiolarians. Migration of siliceous sponges from deep to peritidal environments could be due to an increased amount of nutrients available in peritidal settings. One hypothesis is that the Large Igneous Provinces (LIPs) during the Jurassic possible provided an increased amount of silica into the peritidal Jurassic oceans (Ritterbush et al., 2015).
CHAPTER 9: DISCUSSION

9.1 Previous Observations

9.1.1 Jurassic Studies

The observations of Ritterbush et al. (2014, 2015) served as a key stimulus for using chert and the silica cycle for testing hothouse climate effects. Ritterbush et al. (2014, 2015) found conspicuous and abundant siliceous sponge-dominated Hettangian cherts at shelf depths in South America and Nevada. The fact that these cherts were sponge-dominated attests to the importance of the post-HEATT ocean during the Hettangian in several ways:

1. Carbonate production was sharply reduced in the Hettangian, which Ritterbush et al. (2014, 2015) attributed to ocean acidification. This may have improved the opportunity for the siliceous sponges to capitalize on lack of carbonates and thrive on the shelf. Sponge-dominated cherts had previously moved offshore most conspicuously during the Ordovician (Kidder and Tomescu, 2016).

2. Carter and Hori (2005) showed that radiolarian diversity dropped significantly at the Triassic-Jurassic boundary. It is possible that the reduction in radiolarian extraction and use of silica created a relatively larger supply of silica that allowed the sponges to migrate onto a carbonate-poor platform.

3. Explanations as to the success of sponges during the Hettangian include a combination of reduced radiolarian importance and LIP-supplied silica sources. Which of these factors is dominant is currently unknown, and could be investigated further in
future work. Ritterbush et al. (2014, 2015) hypothesize that the eruption of the CAMP LIP introduced new silica into silica cycle, which has yet to be tested.

Ritterbush et al. (2014; 2015) suggested that silicate weathering of the CAMP basalts stimulated a pulse of spiculitic chert beds on earliest Jurassic shelves at least partly via increasing available dissolved ocean silica. The Karoo-Ferrar LIP may have similarly perturbed the Jurassic silica cycle in the Pliensbachian-Toarcian. Radiolarian diversity dropped sharply at this horizon, but a low-diversity assemblage of apparently opportunistic radiolarians thrived in the wake of this possibly HEATT-driven diversity crash (Goričan et al., 2013).

Conspicuous spiculite beds mark some European sections in the Late Pliensbachian, just as radiolarian diversity was plummeting (Jach, 2002). Sponge abundance may have been favored by an increase in dissolved silica made available by reduced radiolarian demand and/or silicate weathering of Karoo-Ferrar LIP rocks. Data available to date favors the former, as the eruptions appear to post-date the spiculites (Caruthers, 2013). Siliceous-sponge reefs appeared briefly, just after the Pliensbachian-Toarcian HEATT conditions abated, possibly reflecting weathering of the Karoo-Ferrar LIP. Similar siliceous sponge reefs are common in the Upper Jurassic, but are rare in older Jurassic strata. Their brief Middle to Upper Toarcian appearance may hint at a HEATT-stimulated preview of a facies that became common in the Upper Jurassic when sea level was higher and warm climate was perhaps more stable than in the Early Jurassic.
9.1.2 Cretaceous Studies

Some work in the Cretaceous gives support for this thesis. Leckie et al. (2002) observed ocean anoxic events around the Aptian-Albian boundary show some interesting relations. Leckie et al. (2002) also show that the radiolarians are sharply reduced at the Aptian-Albian time interval, as well as during other OAEs. As the percentage of radiolarians drops by more than 50%, the percentage of sponge spicules in these samples rises from about 10% to more than 60% of the biogenic component of the sediment. Kidder and Worsley (2010) characterize this time period as a HEATT episode. The relative abundance of planktonic foraminifera is substantially reduced. This could have been caused by a variety of HEATT factors, most likely acidifaction (Erba et al., 2015, Sabatino et al., 2015).

9.2 Discussion of Results

9.2.1 Chert Abundance

The results of this study show that the frequency of chert deposition increased by over 50% in the Middle and Late Jurassic relative to the Early Jurassic, (Figs. 7.1 and 7.2). Reasons for this spike in frequency of chert deposition could be due to the increase of volcanic activity in the Late Jurassic. Current literature only cites three LIP occurrences at the very latest Jurassic (Table 1.1).

The increased frequency of chert deposition indicates an increase of productivity and abundance of dissolved silica in the oceans. Causes for an increase in dissolved silica could possibly be due to enhanced biological productivity caused by an increase in ocean temperature, or increased silicate weathering of rock produced by a LIP eruption.
A sharp increase in chert abundance occurred between the Aalenian and the Bajocian (Fig. 7.2). Contemporaneously, peritidal cherts significantly decreased in abundance (Fig. 7.6). It is possible that productivity was higher in the Late Jurassic, resulting in more chert and more oil. Klemme (1993) suggests that fourteen petroleum systems with Late Jurassic source rocks contain one-quarter of the world’s discovered oil and gas, as opposed to eleven systems with Lower and Middle Jurassic source rocks, which only contain a relatively minor amount of oil and gas. The increase in chert abundance indicates that the three known Late Jurassic sea-floor LIPs (Table 1.1) could have initiated a HEATT episode. The combined volume from these three Tithonian LIPs is 4.9 million cubic kilometers, which is similar to the 5.0 million cubic kilometers estimate for the Karoo-Ferrar LIP.

In addition to the chert records, the Middle and Late Jurassic intervals are marked by an abundance of siliceous sponge reefs in subtidal, distal shelf, and slope settings (Brunton and Dixon, 1994). The abundance of chert during these intervals as well as the abundance of siliceous sponge reefs argue that the silica availability was high in the Middle and especially in the Upper Jurassic. Why there was such an increase in silica abundance is still unknown. One hypothesis is that the silica increase was caused by the eruption of the Tithonian LIPs (Table 1.1). However, the Tithonian LIPs do not help to explain the increase in silica during the Middle Jurassic.

Another explanation for the increase in silica during the Middle Jurassic is that organisms may have evolved increased efficiency in extracting dissolved silica from the oceans. Harwood and Nikolaev (1995) note that the diatoms most likely first appeared in
the Jurassic, and that they may have been only weakly silicified. Harwood and Nikolaev (1995) also suggest that the diatoms may have originated in nearshore and shelf settings, then eventually migrated to more offshore settings. If weakly silicified diatoms were abundant, they perhaps extracted a large amount of dissolved silica from surface waters. As they sank into the facies that lay beneath them in the water column, their silica would likely have dissolved. It is possible that the diatoms helped to focus delivery of dissolved silica towards to the shelves where they lived at the time.

Lower Jurassic chert abundance may be low because radiolarian production was suppressed by the two well-known HEATT events at the Triassic-Jurassic boundary and at the Pleinsbachian-Toarcian. Sponge cherts on the shelf may have thrived because radiolarians were suppressed, quite possibly as a result of the end-Triassic extinction and HEATT episode. If overall radiolarian diversity was low because of weakened ecosystems, shortages of radiolarian silica in the Lower Jurassic may explain why the Middle and Upper abundance is relatively higher. It is possible the radiolarian ecosystems were simply healthier after they had time to recover after the Lower Jurassic.

9.2.2 Depositional Patterns

The change in chert depositional patterns suggests an adjustment of the silica cycle. It is possible that the diatoms could be starting to influence the silica cycle. Harwood and Nikolaev (1995) suggest that diatoms first appeared during the Jurassic. If diatoms indeed began affecting chert deposition and silica availability, presumably most effectively at upwelling zones, their presence might explain why Middle Jurassic siliceous sponge reefs become important on the shelves. The diatoms displaced the
siliceous sponges that were able to thrive in shelf settings into deeper waters. The sudden
decline in peritidal cherts may be diatom related. It is possible that the diatoms used up
too much of the available silica in peritidal settings, forcing the sponges and radiolarians
to migrate to the shelves. Maliva et al. (1989) thought this was important in the early
Cenozoic. Maldonado et al. (1999) suggested that the diatom effect on radiolarian and
sponge migration began in the Cretaceous. It is possible the results of this study indicate
subtle evidence for a Jurassic application of the Maldonado et al.’s (1999) diatom model.

Another explanation of enhanced chert deposition would be the occurrence of a
HEATT episode possibly brought on by a continental LIP. Starkey and Kidder (2014)
linked the Pleinsbachian-Toarcian HEATT episode to shelf chert deposition. Although
HEATT episodes are expected to occur more rapidly than at the stage level, it is widely
accepted that the Triassic-Jurassic HEATT episode was prompted by the Central Atlantic
Magmatic Province (CAMP) LIP, and the Pliensbachian-Toarcian HEATT episode was
triggered by the formation of the Karoo-Ferrar LIP (Fig. 4.1) (Pálfy and Smith, 2000;
Guex et al., 2001; Wignall, 2001; Courtillot and Renne, 2003; Morard et al., 2003;
Caruthers et al., 2013). The timing of the mass extinctions and ocean anoxic events
(OAE) coincide with the time interval that the Jurassic LIPs were erupting (Fig. 3.1)
(Kidder and Worsley, 2010). Widespread black shale deposits throughout Europe show
evidence for OAEs during these intervals (Hesselbo and Pienkowski, 2011). Additionally,
high seawater temperatures are inferred from isotopic analysis (Cohen et al., 2007;
Jenkyns, 2010; Jenkyns et al., 2002; McArthur et al., 2008). These effects documented
from the Hettangian and Pliesbachian-Toarcian HEATT episode might be reflected again during the Tithonian, when there were three sea-floor LIPs (Table 1.1).

Table 7.3 and Figure 7.4 show that bedded cherts were the dominant chert depositional type throughout the Jurassic. Nodular cherts that occurred alongside bedded cherts were equally abundant in the Early and Middle Jurassic, but there is only one occurrence in the Late Jurassic. Jurassic nodular cherts were absent in the chert record until the Late Jurassic, when nodular cherts accounted for 15% of chert occurrences.

Assuming that the numbers of undetermined cherts do not overturn the trends in the depositional categories, cherts were deposited mainly in shelf settings until the Oxfordian when deep marine cherts became more prominent (Table 7.5 and Fig. 7.6), although shelf cherts were still significant. Cherts deposited in peritidal environments rivaled shelf cherts in frequency until the Pliensbachian, when shelf cherts began to drastically outnumber both peritidal and deep water cherts. There is a gap in the peritidal chert record during the Aalenian, where peritidal cherts are completely absent. Reasons as to changes in prominent depositional environments could be due to migration of siliceous organisms due to silica availability. Recent studies suggest that siliceous sponges, and potentially other siliceous organisms such as radiolarians, overtake disturbed carbonate habitats and initiate shifts to alternative ecosystem (Norström et al. 2009). The ocean euxinia and increased water temperatures associated with HEATT episodes would have disrupted populations of many micro-organisms. This would allow siliceous organisms to inhabit shallow water environments due to lack of competition for nutrients, as well as the availability of silica in shallow environments.
The silica flux from the CAMP LIP would provide a global increase in silica supply in addition to regional volcanic silica sources.

Figures 7.7 and 7.8 demonstrate chert fossil types through time. The entirety of Jurassic cherts contain more radiolarian fossils than siliceous sponge spicules (Table 7.9). The highest frequency of cherts with siliceous sponge spicules occurs during the Middle Jurassic.

Table 7.9 shows the significance between fossil types and depositional environments through time. Early Jurassic cherts were predominately radiolarian cherts deposited in shelf environments. Middle Jurassic cherts also originated in shelf settings and were also abundant with radiolarians, however the frequency of these cherts was almost twice that of the Early Jurassic cherts. Late Jurassic cherts shifted to being deposited in mostly deep marine settings. It is evident that there was either an increase in dissolved silica or productivity during the Middle Jurassic to allow for the increase in frequency of chert occurrences. A relatively cooler environment during the Late Jurassic would allow for the radiolarians to migrate to offshore settings where their remains would be more likely to accumulate in deeper water deposits, which could be one possible explanation. Chatalov et al. (2015) inferred cooling episodes during the Late Jurassic. The increase in abundance of radiolarian cherts during the Late Jurassic could be explained by the increased volcanism associated with the three sea-floor LIPs (Table 1.1). The rationale here is that the significant eruptive volume of these LIPs would contribute considerable volumes of dissolved silica and nutrients to ocean waters. This would stimulate phytoplankton productivity, and hence radiolarian abundance. The
effects on ocean biotas during intervals of active sea-floor LIP emplacement as opposed to the effects of on-land LIP activity have yet to be compared. The Jurassic chert record presented herein provides one framework through which to view such differences.

The most common lithologic facies associated with Early Jurassic chert were limestone, shale, and sandstone, respectively (Table 7.10). The average modern latitude of these deposits are within almost 10 degrees of each other at 34°N and 23°N degrees. The frequency of chert being associated with these facies indicate that the depositional environment fluctuated between peritidal and deeper settings throughout the early Jurassic.

The most common Middle Jurassic cherts are currently located slightly north of Early Jurassic cherts (Table 7.11). The most common associated facies were limestone, volcanics, and shale, which are currently located within 10 degrees of one other at 34°N and 43°N. The facies associated with cherts in the Middle Jurassic point to a period of frequent volcanic activity with the depositional environment of chert fluctuating between shelf and deep marine settings. Cather et al. (2009) suggested volcanic activity during the entirety of the Middle Jurassic.

Late Jurassic cherts were most commonly associated with limestone and volcanics facies (Table 7.12). Present day latitudes of these cherts are located at 34°N and 35°N degrees latitude on average. It is possible that the most common Late Jurassic chert associated facies help to confirm the findings that the most common depositional environments during the Late Jurassic were deep and shelf settings.
Both the Early and Middle Jurassic were most abundant in red and gray cherts (Tables 7.13 and 7.14). The most frequent chert colors during the Late Jurassic shifted to red and green cherts. The consistency of red cherts being most common regardless of age or depositional environment suggests that perhaps chert color is unrelated to relative depositional environment (Table 7.15). An alternative is that the abundance of red chert may be one line of weak evidence suggesting that Jurassic oceans were often oxidized. Some areas and/or intervals were more reduced (e.g. gray and green cherts), and others were strongly reduced to anoxic.
CHAPTER 10: CONCLUSIONS

A globe-encompassing assessment of the spatial and temporal distribution of Jurassic bedded and nodular cherts was attempted in order to confirm or deny the following four hypotheses: 1) Large Igneous Provinces (LIP) stimulated sponge-generated shelf (or peritidal) cherts during HEATT episodes. 2) Sea-floor LIP activity at the latest Jurassic may stimulate biogenic silica production by introducing new silica and nutrients into the ocean. 3) The rise of the diatoms in the Jurassic may stimulate chert both in terms of abundance and nearshore settings.

The conclusions of the study are as follows:

1. Results show that the earliest Jurassic cherts were most dominantly deposited in shelf and peritidal settings. This finding challenges the conventional thinking of most Jurassic cherts originating in deep water settings. It is reasonable to suggest that the Central Atlantic Magmatic Province (CAMP) LIP provided sufficient silica to the oceans to generate shelf and peritidal chert deposits as hypothesized by Ritterbush et al. (2015). Figure 7.6 demonstrates that the Pleinsbachian-Toarcian HEATT episode caused in part by the Karoo-Ferrar LIP is consistent with a relative increase in shelf-originated chert deposits relative to previous periods. Radiolarian cherts were more abundant than expected, indicating that some radiolarian species were able to thrive during HEATT conditions. This case appears to be an example of an instance in which the deposits of a rapid HEATT episode are abundant and conspicuous enough in the record to appear in stage-level tabulations.
2. Hypothesis 2 is supported by Figure 7.6, which exhibits significant chert deposition at the end Triassic-early Jurassic and Pliensbachian-Toarcian HEATT episodes. However, Fig. 7.6 also shows increased chert deposition during the Middle and Late Jurassic. The Late Jurassic is marked by three sea-floor LIPs that emitted a combined basalt volume that is nearly equivalent to the Pliensbachian-Toarcian Karoo-Ferrar LIP, so the combined effect of these three LIPs may provide a window as to the effects. Less encouragingly, the Middle Jurassic and older parts of the Upper Jurassic (Oxfordian-Kimmeridgian) do not have any significant LIP activity.

3. That the Aalenian through Tithonian are particularly chert rich might be the effect of increasing influence of diatoms on the Jurassic silica cycle. Unfortunately, the diatom record is so poor that this possibility is difficult to test. Alternatively, increased silicic LIP volcanism (SLIP) suggested by Cather et al. (2009) could help explain increased Middle Jurassic chert abundance. However, the duration of the SLIP activity begins before the Middle Jurassic and ends before the exceptionally chert-rich latest Jurassic.
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Appendix: Table of Jurassic Chert Compilation

Appendix Table. Compilation of cherts in this table are grouped by categories noted along the top. Depositional Environments were often interpreted by the authors reporting, particularly in sources published within the last 15-20 years. Some depositional environment interpretations are based on facies associations as explained in the Methods section. Abbreviations for the “Chert type” category include: B = Bedded, N = Nodular. Abbreviations for the “Fossil Type” category refer to the silica secreting organisms present in the chert. Abbreviations include: R = Radiolarian, SS = Siliceous sponge.

<table>
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<th>#</th>
<th>Country</th>
<th>Locality</th>
<th>Lat./Long.</th>
<th>Epoch</th>
<th>Stage</th>
<th>Name</th>
<th>Associated Facies</th>
<th>Color</th>
<th>Depositional Environment</th>
<th>Chert Type</th>
<th>Fossil Type</th>
<th>References</th>
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<td>Late Bajocian-Early Bathonian, Middle Bathonian-Early Callovian, Middle Callovian-Early Oxfordian</td>
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<td>Basaltic lavas, breccias</td>
<td>Undet.</td>
<td>Undet.</td>
<td>B</td>
<td>R</td>
<td>Robertson and Shallo, 2000</td>
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<td>Bosnia</td>
<td>Jezeračka Reka</td>
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<td>Undet.</td>
<td>Limestone</td>
<td>Undet.</td>
<td>Shelf</td>
<td>B</td>
<td>SS</td>
<td>Bragin et al. 2011</td>
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<td></td>
<td>Maslovare-Teslic</td>
<td>44°25'49.8&quot;N, 17°37'52.0&quot;E</td>
<td>Middle</td>
<td>Late Bathonian-Early Callovian</td>
<td>Undet.</td>
<td>Shale, limestone</td>
<td>Undet.</td>
<td>Shelf</td>
<td>B</td>
<td>SS</td>
<td>Bragin et al. 2011</td>
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<td><strong>Country</strong></td>
<td><strong>Region</strong></td>
<td><strong>Latitude/Longitude</strong></td>
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<td>Alenian-Cretaceous Xialu Chert Siliceous and tuffaceous mudstone Red Deep B R</td>
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<td>29°13'50.3&quot;N, 88°52'06.5&quot;E</td>
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<td>Kimberidgian-lower Tithonian Undet. Volcanic facies Red, green Deep B R</td>
<td>Baxter et al., 2011</td>
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<td>Lower Callovian-Cretaceous Undet. Punta Conchal Formation Basalt, limestone Red, brown Deep B R</td>
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<td>Undet. Undet. Shale Red, gray, green Shelf B R</td>
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<td>Tithonian Portland Limestone, siliceous sponge spiculites Dark gray Undet. N SS</td>
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<td>Western Sicily</td>
<td>Middle-Late</td>
<td>Early Bajocian-Middle Tithonian</td>
<td>Monte Genuardo</td>
<td>Undet.</td>
<td>Green, gray, red Deep B R</td>
<td>Chiari et al., 2008</td>
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<td>29</td>
<td>Southern Alps</td>
<td>Undet.</td>
<td>Middle</td>
<td>Early Bajocian</td>
<td>Trento Plateau</td>
<td>Undet.</td>
<td>Green, gray, red Peritidal B SS</td>
<td>Woodfine et al., 2008</td>
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<td>30</td>
<td>Central</td>
<td>42°59'59.5&quot;N, 13°00'00.0&quot;E</td>
<td>Early</td>
<td>Toarcian</td>
<td>Rosso Ammonitico</td>
<td>Limestone</td>
<td>Undet.</td>
<td>Peritidal B R</td>
<td>Gill et al., 2004</td>
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<td>32</td>
<td>Japan</td>
<td>Central</td>
<td>Early</td>
<td>TJ-Toarcian</td>
<td>Tamba-Mino Belt</td>
<td>Claystone, siliceous mudstone, sandstone</td>
<td>Red, green, black, gray Shelf B R</td>
<td>Ishiga and Dozen, 1997</td>
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<td>Southwestern</td>
<td>35°32'22.8&quot;N,136°35'25.1&quot;E</td>
<td>Early TJ-Toarcian Mino Complex Shale Black, white, green</td>
<td>Peritidal B R</td>
<td>Hori, 1997</td>
<td></td>
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<td>34</td>
<td>Northwestern</td>
<td>38°57'37.8&quot;N,140°15'58.3&quot;E</td>
<td>Middle Undet. Sandstone, mudstone Undet.</td>
<td>Deep B R, SS</td>
<td>Izozaki et al., 1990</td>
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<td>35</td>
<td>Southwestern</td>
<td>24°22'12.9&quot;N,124°08'25.8&quot;E</td>
<td>Early Late Pleinsbachian-Early Toarcian Fu-saki Formation Mudstone, sandstone, limestone, mafic rocks Gray, red</td>
<td>Shelf B R</td>
<td>Nakae, 2013</td>
<td></td>
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<td>36</td>
<td>Central</td>
<td>33°32'24.4&quot;N,13°15'43.7&quot;E</td>
<td>Early Undet. Ino Formation Mudstone Gray, red</td>
<td>Peritidal B R</td>
<td>Hori and Wakita, 2004</td>
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<td>Central</td>
<td>35°23'56.2&quot;N,16°56'04.1&quot;E</td>
<td>Early Hettangian-Toarcian Undet. Undet.</td>
<td>Red, black, gray</td>
<td>Sato et al., 2012</td>
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<td>40</td>
<td>Nepal</td>
<td>Western Himalayas</td>
<td>27°55'28.1&quot;N,83°49'55.2&quot;E Late Upper Callovian-Tithonian Dras Unit Pillow lavas</td>
<td>Deep B R</td>
<td>Sato et al., 2012</td>
<td></td>
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<td>43</td>
<td>Oman</td>
<td>Western</td>
<td>23°17'46.0&quot;N,55°37'11.1&quot;E Late Undet. Hajar Supergroup Limestone</td>
<td>Undet. Undet.</td>
<td>Kazmin et al., 1986</td>
<td></td>
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<td>Calamian Islands</td>
<td>12°03'15.5&quot;N,12°0'10.31.1&quot;E</td>
<td>Middle-Late Bajocian-Tithonian Laminang cong Formation Siliceous mudstone</td>
<td>Undet. Shelf?/Deep B</td>
<td>Sato et al., 2012</td>
<td></td>
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<td>Poland</td>
<td>Southeastern</td>
<td>52°13'28.2&quot;N,20°47'37.0&quot;E Late Upper Oxfordian Go ry Swietokrz Limestone Gray, brown</td>
<td>Peritidal N Undet.</td>
<td>Sharp et al., 2002</td>
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<td>49</td>
<td>Puerto Rico</td>
<td>Southwestern</td>
<td>Middle Late Bajocian-Early Callovian</td>
<td>Mariquita Chert</td>
<td>Melange, peridotite, basalt, amphibolite</td>
<td>Green, gray, white, red</td>
<td>Undet.</td>
<td>B R</td>
<td>Bandini et al., 2011</td>
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<td>Russia</td>
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<td>Khabarovsk Complex Erdagau Unit</td>
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<td>B R</td>
<td>Suzuki et al., 2005</td>
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<td>51</td>
<td>Russia</td>
<td>Southern Sikhote-Alin</td>
<td>Late Late Kimmeridgian-Middle Tithonian</td>
<td>Undet.</td>
<td>Basalts, turbidites</td>
<td>Green</td>
<td>Undet.</td>
<td>B R</td>
<td>Kemkin and Taketani, 2004</td>
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<td>Late Late Kimmeridgian-Cretaceous</td>
<td>Undet.</td>
<td>Khoi Formation</td>
<td>Undet.</td>
<td>Red, green, gray, pink, green</td>
<td>Shelf</td>
<td>B R</td>
<td>Zyabrev, 2011</td>
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<td>53</td>
<td>Russia</td>
<td>Undet.</td>
<td>Late Late Tithonian</td>
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<td>Indus Formation</td>
<td>Basalt</td>
<td>Green, gray, pink, green</td>
<td>Deep</td>
<td>B R</td>
<td>Danelian and Robertson, 1997</td>
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<td>Northeast</td>
<td>Middle-Late Bajocian-Kimmeridgian-Bathonian-Early Tithonian</td>
<td>Undet.</td>
<td>Avala Gora</td>
<td>Melange</td>
<td>Red, brown, gray, green</td>
<td>Deep</td>
<td>B R</td>
<td>Luchitskaya et al., 2005</td>
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<td>Zaboj</td>
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<td>Shelf</td>
<td>B SS</td>
<td>Bragin et al., 2010</td>
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<td>Middle</td>
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<td>Limestone</td>
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<td>Shelf</td>
<td>B Undet.</td>
<td>Bragin et al., 2010</td>
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<td>Mali Rzav River</td>
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<td>Shelf</td>
<td>B Undet.</td>
<td>Bragin et al., 2010</td>
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<td>Southern</td>
<td>36°41'05.1&quot;N, 5°04'45.0&quot;W</td>
<td>Middle-Late Bathonian</td>
<td>Undet. Oolitic limestone</td>
<td>Undet.</td>
<td>Undet.</td>
<td>1998.</td>
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<td>46°40'47.0&quot;N, 9°53'17.4&quot;E</td>
<td>Early Lower Hettangian</td>
<td>Undet.</td>
<td>Limestone</td>
<td>Undet.</td>
<td>B SS</td>
<td>Sulser and Furrer, 2008</td>
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<td>Turkey</td>
<td>Southwestern</td>
<td>39°52'36.1&quot;N, 29°10'02.8&quot;E</td>
<td>Middle Late Bathonian-Early Callovian</td>
<td>Undet. Bornoval Flysch Zone</td>
<td>Mudstone, Volcanic facies</td>
<td>Undet.</td>
<td>Deep B R</td>
<td>Tekin and Gencuoglu, 2009</td>
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<td>73</td>
<td>United States</td>
<td>Southern Alaska</td>
<td>59°17'24.8&quot;N, 151°42'49.3&quot;W</td>
<td>Early TJ-Sinemurian</td>
<td>McHugh Complex</td>
<td>Basalt</td>
<td>Undet.</td>
<td>B N R</td>
<td>Kusky and Bradley, 1999</td>
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<td>Southern California</td>
<td>35°05'44.3&quot;N, 120°17'49.2&quot;W</td>
<td>Late Middle Oxfordian-Late Tithonian</td>
<td>Undet. Mudstone, limestone, basalt</td>
<td>Undet. Black, green, red</td>
<td>Undet.</td>
<td>Undet. R</td>
<td>Hull and Pessagno, 1994</td>
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<td>36°59'50.3&quot;N, 111°08'07.4&quot;W</td>
<td>Early Pleinsbachian-Toarcian</td>
<td>Undet. Navajo Sandstone</td>
<td>Limestone, sandstone</td>
<td>Undet.</td>
<td>B N Undet.</td>
<td>Bryant et al., 2013</td>
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