

STABLE ISOTOPE RECORD OF SOIL CARBONATES FROM THE EOCENE-
OLIGOCENE TRANSITION, BADLANDS NATIONAL PARK,
SOUTH DAKOTA, USA

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ABSTRACT

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Sections in and around Badlands National Park, Custer County, South Dakota contain a wealth of faunal and floral data within the White River Group. The Chadronian and Orellan North American Land Mammal Ages define the Eocene-Oligocene boundary here. Faunal and floral changes associated with the Chadronian-Orellan interval have long been attributed to the global Greenhouse-Icehouse climatic transition.

Stable isotopes were obtained from paleosol carbonates across the Chadronian-Orellan boundary at three locations in Badlands National Park. Results show wide fluctuations of oxygen isotopes in the uppermost Chadronian and lower Orellan, while carbon isotopes remain relatively stable. Results indicate decoupling of the terrestrial climate system from the marine system. Local conditions appear to play an important role in controlling the record of stable isotopes in paleosol carbonates in the region.

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CHAPTER I: INTRODUCTION

The purpose of this research is to understand the effect of major global climate change on the interior, mid-continental United States. Climate records across the globe are extensive throughout the Eocene and Oligocene. It is well known that Eocene warming occurred on a planet-wide scale, affecting the ecology of all regions. Near the end of that geologic epoch temperatures began to slowly decrease and continental ice masses grew, particularly on Antarctica. This slow temperature decline abruptly accelerated at the end of the Eocene, leading to markedly cooler temperatures globally, with higher latitudes experiencing the most severe changes. This shift from warm temperatures to cool temperatures worldwide is often referred to as the "Greenhouse to Icehouse" transition. In fact, the geologic record of the Eocene-Oligocene transition shows such a dramatic "break" in Europe that Cavelier et al. (1981) propose the boundary between the Paleogene and the Neogene be placed there.

The main line of evidence for these trends is documented in marine oxygen isotopes. The climate record is not as well defined in the terrestrial environment. Several researchers have attempted to ascertain the full impact of continental climate change in the United States during this time period using sedimentary deposits, plant and animal

fossil records (Hutchison, 1982; Retallack, 1983; 1992; Evanoff et al., 1992; Prothero, 1994; Bestland et al., 1997; Kraus, 1997; Prothero and Whittlesey, 1998; Evans, 1999; Prothero, 1999; Evans and Welzenbach, 2000; Stoffer, 2003; Zanazzi et al., 2007) at several locations within the country. The Badlands of South Dakota have several exposed stratigraphic sequences from the Eocene-Oligocene that provide for especially good research opportunities. Retallack (1983; 1985; 1998) has identified a series of paleosols created in a fluvial depositional scheme during this time, while Prothero and others (Swisher and Prothero, 1990; Prothero and Swisher, 1992; Prothero and Whittlesey, 1998) have completed magnetostratigraphy and radiogenic dating, and other researchers (Retallack, 1992; Evanoff et al., 1992; Evans, 1999; Prothero, 1999; Evans and Welzenbach, 2000; Zanazzi, 2007) have examined the paleontology or paleoecology of the region. The paleosol type and composition, along with fossil evidence, reveal that the Badlands were also experiencing overall cooling. In 1974, Birkeland stated, “although many workers have used soil features to infer past climates, studies that have backed up such interpretations with quantitative soil data are relatively few. It is hoped that work on this aspect of pedology will be more quantitative in the future”. To date, not much has been accomplished on this front for the Eocene-Oligocene, but that is exactly what this research aims to achieve.

The impact global climate change has on interior environments, including responses by flora and fauna, can be determined if a clear temperature and precipitation record can be obtained from geologic formations deposited during the period of climate change. More recent Holocene records of mid-continent climate change show that even small changes in precipitation had a profound effect on geomorphic responses, such as

dune re-activation (Denniston et al., 2007). It stands to reason that the same impact could occur at the Eocene-Oligocene.

Changes in temperature, aridity, and vegetation are recorded in oxygen and carbon isotopes comprising calcium-carbonate deposits within marine sediments as well as fossil soils of the Badlands National Park. Deep marine and shallow coastal records are well defined. By collecting and analyzing the isotopic record from the Badlands strata, a comparison between the terrestrial realm, the nearby regional Gulf-Coast, and the global climate at large can be made. The goal of this comparison is to reveal if the inland global cooling trend mirrors either of the marine records or if micro-climates and local tectonic events may have more influence, thus changing the record on a sub-scale. Additionally, the results may explain the apparent decoupling of the mammal extinctions from other faunal and floral extinctions during the Chadronian-Orellan North American Land Mammal Age (NALMA).

To this end, paleosol carbonate samples were collected at the Eocene-Oligocene transition at the Badlands National Park, South Dakota. The Badlands region of South Dakota and Nebraska offer extraordinary stratigraphy ranging in age from the Cretaceous through the Tertiary. The park offers an excellent location for studying patterns of fossil-bone accumulation, paleosol features, horizons of calcareous nodules and layers, and local variations in mineralogy and trace and major chemical elements (Retallack, 1983). This area is of particular interest due to its location in the center of the continent. Carbonate was collected within paleosol sequences identified by Retallack (1983) and linked to the geologic timeline via magnetostratigraphy by Prothero and Whittlesey (1998). Carbonate samples were analyzed at the Indiana University Department of

Geology using Mass Spectrometry. Generally, calcite $\delta^{18}\text{O}$ in any deposit increases as temperature decreases or as the mass of continental ice increases (Zachos et al., 2001), while $\delta^{13}\text{C}$ from the Eocene-Oligocene indicates the relative abundance of CO_2 in the atmosphere (Cerling, 1991; Zanazzi et al., 2007). In the marine realm $\delta^{13}\text{C}$ reflects changes in the ratio of the dissolved inorganic carbon of ambient seawater, which reflect changes in the rates of carbon supply and removal from organic and inorganic reservoirs (Zachos et al., 2001), such as sea-floor venting.

In addition to comparing the Badlands isotopic record to near-shore and deep marine sequences, it can be compared to other terrestrial records. Qualitative evidence from the park and surrounding region can be compared to quantitative data, linking temperature and precipitation changes to changes in floral and faunal adaptation. This either lends support for qualitative methods, or identifies areas requiring refinement and further study.

CHAPTER II: GEOLOGIC HISTORY OF THE EOCENE-OLIGOCENE

Global

The late Eocene-early Oligocene Earth was transformed by a number of tectonic processes. Every continent underwent change relating to plate movement; either uplift, subsidence, development, rapid movement, or increased volcanic activity. The increase in volcanism and tectonism coincided with, and was related to a change at 38 Myr in the Earth's pole of rotation (Kennett et al., 1985). During the Eocene, North America was connected to Asia by the Bering land bridge across the Pacific Ocean and to Europe through Greenland and Iceland across the Atlantic Ocean (Briggs, 1987; Ivany et al., 2003). South America and Australia were connected to Antarctica as remnants of Gondwanaland in the southern hemisphere (Ivany et al., 2003), as seen in Figure 1. The distribution of oceans and seas was significantly different from the present. Eustatic sea levels were high; the only circum-global ocean current was around the equator, and cross-latitudinal circulation was restricted; the Tethys sea extended from India to the Americas; and the Australo-Antarctic Gulf circulated warm water between Australia and Antarctica, closed off from the Proto-Pacific Ocean by the Tasmanian land bridge (Exon et al., 2002; Ivany et al., 2003). The lack of land barriers between the world's oceans and seas led to

free exchange of waters and taxa between the Pacific Ocean, Gulf of Mexico, Caribbean Sea, Atlantic Ocean and the Tethys Sea.

“At the end of the Eocene, [a] series of geological events occur[ed] on a worldwide scale, of which the intensity is comparable to the one occurring at the Cretaceous/Tertiary boundary” (Cavelier et al., 1981). Among the many tectonic activities occurring in the lead up to and across this boundary, two events stand out as particularly important: 1) the collision of India with Asia, creating the Tibetan Plateau uplift, and 2) the isolation of Antarctica from the remaining continents (Sloan and Barron, 1992; Retallack, 1992; Exon et al., 2002; Ivany et al., 2003). Other tectonic events occurring at this time included the re-alignment of the Pacific Mid-Ocean Ridge (MOR), the Alpine orogenesis in Europe with the associated closing of the Tethyan sea, opening of the Greenland-Norwegian Sea to the Arctic, the breaching of the Rio Grande Rise, and the Andean and Laramide Orogenies in South and North America, respectively (Cavelier et al., 1981; Sloan and Barron, 1992; Retallack, 1992; Ivany et al., 2003).

Australia/Antarctica/Indian Ocean Region

From the late Cretaceous through the early Tertiary, Antarctica served as a bridge connecting the otherwise disparate continents of South America, Africa and Australia. The land mass connections allowed for faunal migration. Ancestral Australian marsupials, turtles, frogs, lizards, ratite birds, and many insects are thought to have immigrated to Australia from South America along this route in the early Tertiary, with members of the marsupial family Polydolopidae being the first land mammals found in Antarctica (Briggs, 1987). New Zealand and Australia developed a complex faunal

relationship that reflects both the variable distances they had from each other and from Antarctica throughout the Tertiary (Briggs, 1987).

The isolation of Antarctica is frequently singled out as the major driving factor for global climate change at the end of the Eocene. From 70-43 Ma slow separation of Australia and Antarctica occurred (Briggs, 1987; Exon et al., 2002). Faster separation is noted from 43-37 Ma, but the two landmasses were still connected by the Tasmanian Land Bridge (Exon et al., 2002). From 42-35 Ma sediments from two Deep Sea Drilling Project (DSDP) Leg 90 sites in the south-west Pacific, and geologic evidence from New Zealand, show a pronounced increase in local, highly explosive volcanic activity - particularly near the Eocene-Oligocene Boundary (Kennett et al., 1985).

The change at 38 Ma in the Earth's pole of rotation caused the development of an active tectonic zone through New Zealand. As the South Pacific Rise and southeast Indian spreading axis merged between Australia and Antarctica an extensional transform boundary across New Zealand formed, effectively representing the development of an Indo-Australian/Pacific plate boundary in the latest Eocene (Kennett et al., 1985). Meanwhile, South America was separating from Antarctica, opening the Drake Passage beginning at 37 Ma (Diester-Haass and Zahn, 1996; Ivany et al., 2003). At 33.5 Ma the continental margin around Tasmania collapsed and subsidence ensued with additional collapses in the Victoria Land Basin east of the rising Trans-Antarctic Mountains and along the coast of southeast Australia; South America was completely separated from Antarctica and the isolation of Antarctica was complete (Diester-Haass and Zahn, 1996; Exon et al., 2002; Ivany et al., 2003). Evidence of these events is in marine sediment cores recovered from Ocean Drilling Project (ODP) site 1182 and Leg 189 in the

Australian-Antarctic seaway. Results indicate a switch from early-rift clastic deposition to carbonate deposition in the late Eocene under deepening waters from 100-500 meters (Lear et al., 2000; Exon et al., 2002). A 20-m thick nannofossil chalk bed was recovered from the earliest Oligocene, suggesting rapidly deepening water depths exceeding 1,000 meters (Kennett et al., 1985; Lear et al., 2000; Exon et al., 2002). Additionally, Indian Ocean spreading rates increased on the south-east Indian Ridge and major changes in spreading directions occurred on both the south-west Indian Ocean and the Central Indian Ridges (Kennett et al., 1985).

South America/Pacific Ocean

The pulse of volcanic activity during the late Eocene-early Oligocene in New Zealand occurred throughout the circum-pacific region. A change in rotation of the Pacific plate, indicated by the bend and increased volume of erupted material in the Hawaiian-Emperor volcanic chain at ~42 Ma, led to pacific-wide increased volcanism associated with large-scale late Eocene plate-tectonic changes (Kennett et al., 1985). Examples include: active marginal basin development throughout the Pacific Ocean, overthrusting of New Caledonia and the formation of new island-arc systems in the south-west Pacific; sudden volcanicity in Japan from 41-36 million years related to the termination of the Shimanto Orogeny and the onset of the Mizuho tectonic phase in north-east Japan; the initiation of the Philippine, Bonin, Mariana, Yap, Palau and Tonga trench-arc systems during the late Eocene; the Mendocino Fracture Zone directional change between the Pacific and Farallon Plates; and increased volcanism in the Peruvian Andes of South America at approximately 40 Ma (Kennett et al., 1985).

Eurasia/Atlantic Ocean

As in other regions, geography had great influence on faunal migrations. From Asia, the flow of fauna was greatest from east to west (Briggs, 1987). Collision between India and Eurasia commenced about 40 Ma (Kennett et al., 1985). The uplift of the Himalayas on the north border of India created a land barrier to all but the montane species. Entry into India was largely restricted to two other gateways-the Assam region to the east and another gateway through the northwest (Briggs, 1987).

At various times throughout the Tertiary, Africa and Eurasia connected across the Tethys Sea (Briggs, 1987). During the Paleocene and Eocene, at least three primitive mammalian groups reached Africa from the north (Briggs, 1987). The bridge Antarctica formed between South America and Africa/Eurasia was an especially important passageway. Fresh water turtles made their way from Asia in the Eocene to currently reside in South America, Africa and Madagascar; while beetles from the Eocene Baltic region are now distributed South America, South Africa and Australia (Briggs, 1987).

From 42-38 million years, intense orogeny is noted in the Illyrian orogenic phase in the North Atlantic and Mediterranean regions (Kennett et al., 1985). European tectonism increased during the late Eocene to the early Oligocene as the strain fields in the Alpino-Mediterranean collision zones re-arranged from NW-SE in the Eocene, and compressive in nature; to NE-SW in the Oligocene, and tending towards relaxation (Cavelier et al., 1981). This led to penetration and folding in the Tethyan region, while the overlying rocks were swept west into the Franco-Italian Alps, beginning the supply of flysch to the external zones of the Alpine chains (Cavelier et al., 1981). As the Tethys

closed at the Eocene-Oligocene boundary, the lateral deep-water passage between the Tethys Sea and the Atlantic Ocean also closed (Cavelier et al., 1981).

Europe was connected to North America via two land bridges in the North Atlantic during the early Eocene -the Thulian route extended from Labrador and Baffin Island through Greenland, the Faeroe Islands and Scotland , and the DeGeer route connected Ellesmere Island, Greenland, Spitsbergen and Scandinavia (Briggs, 1987). The separation of Greenland from Europe began in the early Paleocene (Briggs, 1987) and continued into the Oligocene (Eldholm and Thiede, 1980). Increased volcanic activity at the Atlantic mid-ocean ridge during pole re-alignment is evidenced by a decrease in Manganese in the sediments of the North Atlantic from the Eocene into the Oligocene (Cavelier et al., 1981).

North America/Caribbean

While the connection to Europe across the Atlantic Ocean began separating in the early Eocene, the Beringian region remained connected throughout the Tertiary (Briggs, 1987). North America was essentially divided into two large islands at the beginning of the Eocene, as the Rocky Mountains were separated longitudinally from the rest of the continent by an inter-continental seaway. Florida and much of Central America were submerged, and the northeast-flowing Suwannee Current separated Florida from the Gulf Coast during the Eocene, preventing Tethyan fauna from crossing over to the central Gulf until sea level fell at the Eocene-Oligocene, breaching the divide and leading to a fossil record with Tethyan fauna in western Alabama and Mississippi near the Eocene-Oligocene boundary (Ivany et al., 2003). Tectonic uplift also began occurring in the

Caribbean during the Eocene-Oligocene transition, restricting the Gulf Coast from the Caribbean Sea (Ivany, Nesbitt and Prothero, 2003).

The origin of the Rocky Mountains was long before Eocene-Oligocene, approximately seventy million years ago (Stoffer, 2003). This orogeny created many basins and intervening uplifts in a relatively short amount of time, including the Black Hills in South Dakota (Stoffer, 2003). As the Rockies rose, their subsequent erosion and the unroofing of the coincident Black Hills caused an influx of sedimentation towards the Midwest, which shrank the interior seaway and caused land to rise above sea level (Stoffer, 2003). Un-roofing of the Black Hills was greatest during the early Eocene (Evans and Welzenbach, 2000). The seaway shrank from west to east, creating the floodplain deposits of the Hell Creek Formation in northwestern South Dakota, Montana and North Dakota (Stoffer, 2003).

The Williston Basin in North Dakota and Montana, and the Powder River and Big Horn Basins in Wyoming are the result of the aforementioned uplift, extension and erosional fill during the Eocene-Oligocene (Kraus, 1997; Stoffer, 2003). 700 meters of Eocene sediment comprise the Willwood Formation in the Big Horn Basin, demonstrating the amount of sediment backfill caused by erosion of the rising mountains (Kraus, 1997). The Willwood consists of sheet sandstones deposited by major meandering rivers, overbank deposits and avulsion deposits (Kraus, 1997). Multiple thin (<1 meter) compound paleosols formed on the avulsion and levee deposits (Kraus, 1997).

In central Oregon the Eocene-Oligocene transition is recorded in the Clarno and John Day Formations. The Clarno Red Beds are the Eocene phase of Clarno volcanism with the Upper Red Beds composed of paleosols representing 600,000 to 1.5 million

years of soil formation (Bestland et al., 1997). The John Day Formation, consisting of the Big Basin and Turtle Cove members, represents the onset of Eocene Cascadian Volcanism approximately 40-42 million years ago (Bestland et al., 1997). A thick section of the Big Basin Member is preserved in the Sutton or Mitchell Mountain Syncline of Painted Hills Area (Bestland et al., 1997). The Middle Big Basin truncates the Lower Big Basin member to form the Eocene-Oligocene boundary at approximately 33 Ma (Bestland et al., 1997). The paleosol changes from kaolinite and Fe-rich to Smectite rich and tufaceous, and it is extensively exposed at the Painted Hills area as well as other locations (Bestland et al., 1997).

By the Oligocene, abyssal waters originating in Antarctica began arriving in the North Pacific allowing the Pelagic Meiji body of sediments on the Northern Flank of the Emperor Sea Mounts to begin accumulating (Ivany et al., 2003).

The White River Badlands

In addition to the depositional environments described above, the White River Badlands of South Dakota offer remarkable views of Eocene-Oligocene stratigraphy. The Badlands sit just east of the Black Hills and record a nearly continuous sequence of sediment deposition from the initial inland sea into the Miocene. The age of the Eocene-Oligocene transition in the Badlands National Park is 33.7 Ma (Stoffer, 2003). The Badlands region of South Dakota experienced early Cenozoic uplift and un-roofing of the Black Hills, with incision and backfill in adjacent areas (Evans and Welzenbach, 2000). Faults throughout the Badlands National Park in South Dakota are positioned in a way

that suggests that the Sage Creek Arch Anticline was in progress before the late Eocene and through the Oligocene into the Miocene (Stoffer, 2003).

The sediment stratigraphy of the Badlands cannot be discussed without discussing the paleosol sequences. The late Eocene-early Oligocene White River and Lower Arikaree groups of the park have a long history of controversy over their likely paleosol development (Retallack, 1983). Over 150 years of research in the Badlands region has led to almost every non-marine paleoenvironmental consideration, with most now agreeing that the deposits are an extensive alluvial plain (Retallack, 1983), with mineralogical and chemical evidence of paleosols. The Badlands paleosols comprise the sample set for the isotopic record derived in this study. Therefore, an in depth examination of their identification and paleoecological assessment is discussed below.

Paleosols Retallack sampled showed surficial depletion of potassium oxide (K_2O) and phosphorous pentoxide (P_2O_5) in proportion to their degree of development, a condition commonly found where plants have grown (Retallack, 1983). They also showed an abundance of clay and other minerals along with chemical variations and trace elements within the profile comparable with the degree and kind of development seen in modern analogs; for example, all have a higher and sharper 15\AA smectite peak in their upper horizons (Retallack, 1983).

Parent material of the paleosols was varied and included volcanic ash fall, resorted soil material and distantly derived alluvium (Retallack, 1983). While parent material had a systematic bias on the chemical analysis and molecular weathering, soil formation effects overprinted these (Retallack, 1983). Some examples of the soil formation effects include lessivage (the soil forming process in which the B horizon is

enriched in sesquioxides and clay that have washed out of the A and accounts for the color banding in the Badlands), sepic plasmic fabrics found in thin sections, and the concentration of trace fossils including burrow-like irregularities on the tops of many layers (Retallack, 1983).

Retallack successfully identified paleosol series using the soil development schema of Birkeland, 1974, and applied several soil classification approaches to paleosol series in the Badlands National Park (1983). These included the soil classification method of The US Department of Agriculture Soil Survey Staff of 1975, the Australian Commonwealth Scientific and Industrial Research Organization, and the non-genetic key of Northcote, 1974. 87 successive paleosols were identified in 143 meters of stratigraphic section, including at least ten paleosol series in the Pinnacles area (Retallack, 1983; Stoffer, 2003).

Those paleosol series, in order of stratigraphic appearance, are: 1) Yellow Mounds series: thick, very strongly developed, sandy, yellow-red; 2) Interior series: thick, very strongly developed, clayey, grey-red; 3) Gleska series: strongly developed, clayey, grey-pink; 4) Ohaka series: very weakly developed, clayey, grey-green; 5) Conata series: weakly developed, silty, tan; 6) Zisa series: very weakly developed, silty, red; 7) Ogi series: weakly developed, silty, orange, with iron-manganese haloed root traces; 8) Wisangie series: moderately developed, ashy, light colored, caliche nodules, large drab-haloed root traces; 9) Pinnacles series: moderately developed, ashy, light colored, caliche nodules, fine root traces and 10) Samna series: moderately developed, ashy, dark brown, caliche nodules (Retallack, 1983). In contrast to the kaolinitic Yellow Mounds silty clay loam, White River and lower Arikaree groups are smectitic (Retallack, 1983). The

paleosol series are depicted in Figure 2. They developed amid the sedimentary depositional sequence shown in Figure 3 and herein described.

Pre-Cambrian basement rocks underlie Paleozoic and Mesozoic marine sedimentary layers that terminate in the late Cretaceous Pierre Shale (Evans and Welzenbach, 2000). The Yellow Mounds are the result of Eocene meteoric weathering of the Pierre Shale, producing yellow paleosols forming rolling, hilly terrain of moderate relief (Retallack, 1983; Evans and Welzenbach, 2000; Stoffer, 2003).

Incising and backfill during the early Cenozoic led to 250 meters of non-marine sediments accumulating in the Eocene-Oligocene White River Group east of the Black Hills (Evans and Welzenbach, 2000; Stoffer, 2003). The White River and Arikaree Groups have 3 main kinds of parent material: 1) volcanic ash fall, 2) resorted soil material, and 3) far transported alluvium (Retallack, 1983). “These are mixed in different proportions that are complexly interrelated with the paleoenvironmental changes” (Retallack, 1983). The resulting deposits include sandstones, tufaceous sandstones, conglomerate, mud-rocks (shale and clay mudstone), bone beds, and volcanic ash deposits on floodplains fed by uplifts throughout the Rocky Mountain Region (Stoffer, 2003). From the mid to late Tertiary sediment from the Black Hills and ash fall from the Western Cordilleran region of Nevada began filling in the low valleys; the region grew progressively higher and drier, with river channels and associated floodplains forming in addition to soils from the sediment influx (Stoffer, 2003). Volcanic activity in Nevada continued in the Oligocene, blanketing the Badlands with several ash deposits (Stoffer, 2003). The sedimentation record in the Badlands ends in the Miocene, approximately 20 million years ago (Stoffer, 2003).

The White River Group is composed of the Chamberlain Pass Formation, The Chadron Formation and the Brule Formation (Stoffer, 2003). The Chamberlain Pass and Chadron Formations were deposited in the Eocene, while the Brule makes up the bulk of the Oligocene deposits in the park. The Chadron Formation includes the Interior Paleosols and lacustrine limestones, and can be further subdivided into three members: the Peanut Peak, the Crazy Johnson, and the Ahearn (Stoffer, 2003). The Brule is comprised of the Scenic and Poleslide Members (Stoffer, 2003). General features of the White River Group include a popcorn-weathered surface, with subsurface deposits that are extensively cracked, veined and slickensided (Retallack, 1983). In many cases those features are the remains of original natural aggregates of soil and clay skins and modified surfaces, voids, and cracks (Retallack, 1983).

The Chamberlain Pass Formation is composed of beds that crop out along an unconformable surface between the Pierre Shale and beneath the base of the Chadron Formation, with the upper surface having an Interior Paleosol overprint (Stoffer, 2003).

The Chadron formation ranges from 37-33.9 million years, and is almost entirely a fluvial depositional environment containing several varieties of carbonates (Swisher and Prothero, 1990; Evans and Welzenbach, 2000). It can generally be considered a poor consolidation of mud rocks having a frothy popcorn texture and forming a bench-like surface with rolling incised topography, often called haystacks (Retallack, 1992; Stoffer, 2003). The subtle pink and green color banding reflects a large number of successive buried soil horizons (Retallack, 1992). At its base is a grey to olive-grey, sandy clay with local conglomerate and white, freshwater limestone lenses (Stoffer, 2003). The

conglomerate is composed of chert, granitic rocks, and metamorphic materials from streams draining the Black Hills region or further (Stoffer, 2003).

The grey floodplain sedimentation was likely continuous and slow from the beginning of the seaway withdrawal to the influx of the Laramide sediments, although other theories suggest there was a time gap between these events (Stoffer, 2003). Stream base levels were influenced by global eustatic cycles with backfilling occurring during high sea-level stands and soil weathering profile forming when sea-level and stream base fell (Stoffer, 2003). By the late Eocene, 2.1-2.5 km of sedimentary rocks had eroded, exposing Paleozoic carbonates and Precambrian crystalline rocks (Evans and Welzenbach, 2000). The Paleozoic carbonates were karstified in late Eocene, resulting in carbonate-rich groundwater, which flowed along structural features and discharged into the Chadron depositional environment resulting in the precipitation of tufas, lacustrine limestones and travertines near springs and in fluvial and lacustrine systems (Evans and Welzenbach, 2000). The carbonates are inter-stratified with rocks in the upper part of Chronozone C13R, close to the Chron C13R-C13N boundary southwest of the Badlands National Park (Evans and Welzenbach, 2000). Vertically stacked carbonate deposits are mostly found in fault zones, fracture systems, or along the trend of paleovalleys because they are linked to paleogroundwater flow and discharge at springs and seeps (Evans and Welzenbach, 2000).

Limestones are evaluated as lacustrine tufas because of their close spatial relationships with fluvial tufas, weakly developed shorelines, lack of siliciclastic detritus, presence of encrusted aquatic macrophytes, presence of spring resurgences, tufa breccias and presence of freshwater fossils (Evans and Welzenbach, 2000; Stoffer, 2003). In

contrast to the relationship between lacustrine and fluvial limestones, groundwater fed systems have no relationship with adjacent fluvial systems (Evans and Welzenbach, 2000). The different type of calcareous deposits is important to note. Limestones and other non-pedogenic deposits were not sampled in this study, as the effects of residency in groundwater or spring fed ponds could not be controlled for. Instead, only those calcareous deposits coeval with soil development were sampled.

Two pedogenic calcareous horizons were identified in the Chadron Formation: the first was soft powdery layers in the Chadron Formation with calcic horizons; the second was thin, purple stained, partly silicified, calcareous stringers of paleosols in the upper Chadron Formation and Scenic Member of the Brule formation (Retallack, 1983). These were probably indurated calcareous pans, like petrocalcic horizons of some modern arid land soils (Retallack, 1983).

The sediments in the Badlands sequence preserve bones, traces of invertebrate remains, plant materials, root traces, animal burrows, coprolites, animal tracks, fossil soils, stream and flood plain deposits, and deposits from warm, shallow lakes (Stoffer, 2003). The land snail, *Pseudolisinoc leidyi*, is the most common and widespread fossil found, and is also the only fossil invertebrate found in the Pinnacles (Retallack, 1983). An abundance of bones from large Rhinoceros-like mammals were observed in the Chadron Formation and mapped by early investigators, resulting in the Formation originally being called the Titanotherium beds (Stoffer, 2003). Alligators, frogs, fish, aquatic turtles, bivalve shell fragments, ostracodes, calcified aquatic roots, charophytes, algal filaments, lizards, abundant land tortoises, birds, opossum, small insectivores, rodents, saber-tooth cats, tapiroids, peccaries, camels, and squirrel-like, deer-like,

rabbit-like, and fox-like animals were also found (Retallack, 1983; Evans and Welzenbach, 2000; Stoffer, 2003).

The upper formations of the White River Group are the Brule, which begins in the Oligocene; followed by the Sharps formation (Stoffer, 2003). The 140-meter thick Brule formation consists of the lower Scenic Member, and upper Poleslide member (Evans and Welzenbach, 2000). The Brule Formation consists of fluvial channel and overbank deposits with inter-stratified air-fall tuffs and paleosols with an abrupt reduction of carbonate production compared to the Chadron (Evans and Welzenbach, 2000). Within the Brule are non-pedogenic nodular zones formed of fluvial sheet sandstones cemented after burial and later jointed (Evans and Welzenbach, 2000).

The Scenic member is a series of paleosols on a broad, aggrading floodplain (Stoffer, 2003). The Scenic member alternates rusty red and grayish white, and contains a fairly even mix of ash fall, assorted soil material, and far transported alluvium (Retallack, 1983; Stoffer, 2003). The white indicates stream channel bar and over-bank splay sand, while the red are paleosols formed slowly on a floodplain (Stoffer, 2003). A third calcareous horizon identified in the White River Group is rounded calcareous nodules of paleosols in the Sharps and Poleslide members (Retallack, 1983). These are similar to the calcite of modern soils (Retallack, 1983).

Petrographic examination of three Scenic member paleosols (Gleska, Conata and Zisa) reveal that in the Gleska and Conata paleosols, there was some illuviation of clay from the A horizon-indicated by clay skins (Retallack, 1983). Clay enrichment of the B-horizon is most pronounced in the Gleska, less pronounced in the Conata, and absent in the Zisa (Retallack, 1983). Additionally, the Gleska have oxidized B-horizons and drab

C-horizons (Retallack, 1983). The water table was probably within one meter of the surface (Retallack, 1983).

Scenic Member deposition marks the Chadronian-Orellan NALMA transition. The Chadronian-Orellan boundary is presented as an unconformity in some parts, and as non-fossil bearing rocks in other parts of the Badlands (Prothero and Swisher, 1992). Truncated ranges of mammalian fossils at the base of the Scenic member may indicate a major disconformity, which is supported where the layer also contains relict petrocalcic horizons (Retallack, 1983). However, deposition of the Chadron into the Scenic Member is basically continuous, as the Scenic Member conformably overlies the Chadron in most locations despite a notable unconformity in the Conata Picnic area (Prothero and Whittlesey, 1998; Stoffer, 2003).

Swisher and Prothero (1992) utilized the advanced Argon (Ar) isotopic dating method, $^{40}\text{Ar}/^{39}\text{Ar}$, and more thorough sampling techniques than previous studies to place the Chadronian-Orellan NALMA at 33.9 Ma in the Badlands National Park (1992). Whittlesey added to the Badlands magnetostratigraphy coverage with sections in the type Scenic area (Chamberlain Pass) and in the Conata Basin, near the Pig Dig (Prothero and Whittlesey, 1998) (Figure 4). When aligned with the magnetostratigraphy for the region, the White River Group lies within Chron C13R, at the Eocene-Oligocene transition (Swisher and Prothero, 1990). Early Orellan normal polarity is very short, occurring just below or at the base of the "lower nodules," while the reversal stretches from the Scenic Member through the lower half of the Whitneyan Section of the Poleslide Member (Prothero and Swisher, 1992). Additional support for this chronology comes from Flagstaff Rim, Wyoming (Prothero and Swisher, 1992). The middle and upper part of the

Flagstaff Rim section can be correlated with Chron C13R based on Swisher and Prothero dates (Prothero and Swisher, 1992).

Fossils found in the Chadron Formation and Scenic Member of the Brule include aquatic turtles (*Graptemys*), semi-aquatic rhinos (*Trigonias*), cursorial rhinos (*Hyracodon*, *Caenopus*), titanotheres (*Menodus*), horses (*Mesohippus*), creodont carnivores (*Hyaenodon*), oreodonts (*Merycoidodon*), and the pig-like enteledon (*Archaeotherium*) (Retallack, 1983; Evans and Welzenbach, 2000). At the Chadronian-Orellan contact, alligators, many turtle species, and the large titanotheres became extinct—only one titanotheres and one small alligator were found in the Brule (Retallack, 1983). Fossil land snails were also reduced in diversity (Evanoff, 1992).

Orellan fossils come mainly from two zones at the Badlands National Park—lower and upper nodules (Prothero and Swisher, 1992). Oreodonts (*Merycoidodon* variety) and turtle beds were found in the lower Scenic Member, while the upper Scenic Member channel deposits were originally called the *Metamyndon* beds after the bones of aquatic rhino (Stoffer, 2003). The horned sheep herbivore, *Proteroceras*, and the oreodont, *Leptauchenia* are found in the Poleslide member (Stoffer, 2003).

Preservation biases in soil types must be assessed, as some fossil soils preserve more than others. Bone was most abundant in weakly developed, calcareous paleosols, presumably alkaline, but also common where calcareous paleosols were moderately to strongly developed (Retallack, 1983; 1998). Non-calcareous paleosols did not preserve bones, likely because that environment is acidic (Retallack, 1983; 1998).

Conata and Gleska paleosols of the Scenic member contain an abundance of bone and articulated skeletons (Retallack, 1983). In the Gleska of the Chadron, almost no bone

was found, because they were not alkaline enough for preservation (Retallack, 1983). Fossil mammal bones were usually poorly preserved and disarticulated when found in the paleochannels and associated sediments, which lead to an extremely biased mammal population sample (Retallack, 1983). Where preservation is exceptional, such as in the Scenic Member, seasonal migrations and size related preservational biases complicate exact reconstructions of mammalian populations, and only a general impression of communities of large mammals is attainable (Retallack, 1983). Temporal gaps at the Badlands make it a bad place to study mammal fossils (Retallack, 1998).

The upper Poleslide Member is a light grey massive cliff forming sandstone which formed by larger stream channels in drier conditions than the Scenic Member (Stoffer, 2003). Overlying the Poleslide is the sandstone and mud rock of the Sharps Formation (Stoffer, 2003). The Sharps formation is 26-28 million years old, making it the latest Oligocene strata in the Arikaree Group (Stoffer, 2003). It forms the highest peaks, and at its base is the Rockyford Ash (Stoffer, 2003). The Sharps Formation is composed of ignimbrite volcanics, tuffaceous sandstone, stream channel sand and floodplain mud paleosols of drier steppe or desert conditions (Stoffer, 2003). The parent material of the Poleslide and Sharps formations of the Arikaree Group is mostly ash (Retallack, 1983).

As nearby tectonic uplift and unroofing with coeval incising and backfill continued, the paleotopography, and hence geomorphology and groundwater elevations, were continuously altered. During the early Tertiary, the elevation was near sea-level, but began to rise by the mid-Tertiary (Stoffer, 2003). The water table in the early Oligocene Gleska series was probably within 1 meter of the surface, and the sinuous paleochannels of the Chadron formation were no more than 3 meters deep (Retallack, 1983). Cut banks

shallower than 2 meters lead to broad, low-gradient flood plain topography (Retallack, 1983). The early to late Oligocene may have had a comparably high water table in the Gleska, but there is less evidence of it, most likely because there was a decrease in mean annual rainfall (Retallack, 1983). In the late Oligocene groundwater depths were decreasing, as evidenced by depths to calcareous nodular horizons at Pinnacles declining higher in the section (Retallack, 1992). Paleochannels incised 1-2 meters deep and were loosely sinuous, with some braiding (Retallack, 1983). This may be due to sparser vegetation, but is likely due to low local topographic relief (Retallack, 1983). Continued low topography is indicated by more sinuous, shallow (less than or equal to 2 meters) incising paleochannels of the Poleslide (Retallack, 1983). The late Oligocene Sharps formation showed some evidence of surficial water logging in streamside Ogi series, and some Samna series, but evidence of a high water table is lacking (Retallack, 1983).

Using the paleomagnetic and radiometric dating in the Badlands National Park, sediment accumulation rates for the Pinnacles can be calculated (Retallack, 1983). Estimates are low because the sequence is incomplete and likely includes additional disconformities not noted, and because rock units do not correspond exactly with chronostratigraphic units (Retallack, 1983). Sedimentation rate estimates are: Chadron Formation = 0.0047 mm/yr, Scenic Member = 0.023 mm/year and Poleslide Member = 0.027 mm/year (Retallack, 1983). The estimates are averaged, as some increments in the Scenic member accumulated 50 cm of sediment at a time (Retallack, 1983).

Sedimentation accumulation rates can also be estimated from directly observing the paleosols and their level of development. This is useful in units that are not able to be dated (Retallack, 1983). Based on known soil development time frames, the White River

and lower Arikaree Groups in the Pinnacles area potential time of deposition varies from 10^0 to 10^6 years (Retallack, 1983). Additional time may have been unrecorded in paleosols subsequently incorporated within the recognizable paleosol profiles (Retallack, 1983). The timeframe for development does not allow for adequate interpretation of the effect of geologically short events, such as climatic fluctuations, because the time span for Pleistocene glacial and interglacials is 10^3 to 10^5 years (Retallack, 1983).

Nevertheless, sedimentation rates estimated from paleosol development are 0.28 mm/yr for the Chadron Formation, 0.47 mm/yr in the Scenic Member and 0.65 mm/yr in the Poleslide Member (Retallack, 1983). The paleosol development rates are much higher than those estimated from dating, but show comparable changes from one unit to another and can be used to compare individual successions of paleosols (Retallack, 1983). The lower Scenic Member rate is 0.38 mm/yr, while middle and upper Scenic Members are 0.22 mm/yr (Retallack, 1983). Sediment accumulation rates increased in the lower Poleslide Member, 1.06 mm/yr but decreased in the upper Poleslide, 0.50 mm/yr (Retallack, 1983), which is likely due to rates of volcanic ash deposition. Sedimentation rates remained high during the deposition of the Sharps Formation (Retallack, 1983).

As previously mentioned, many Badlands paleosols contour calcareous layers and nodules arranged in horizons with a predictable relationship to the rest of the paleosol (Retallack, 1983). Badlands paleosol calcite layers and nodules are arguably original soil features for the following reasons: there is a lack of aquatic fossils; there is lacustrine bedding or varving within these limestones; there is a greater abundance of weathering materials within the nodules compared to the surrounding matrix; there is observed truncation of calcareous horizons by the cut-bank of sandstone paleochannels; there is a

presence of similar calcareous clasts within intraformational breccias of paleochannels; and there is a continuation of drab haloed root traces through the nodules from the adjacent matrix (Retallack, 1983). Strontium isotopic and petrographic studies of the White River and lower Arikaree Groups show that the widespread calcite and smectite cements appear to have formed at about the same time as the sediments and “remain[ed] in a chemical equilibrium with modern groundwater” (Retallack, 1983). Late diagenetic alteration was confined to local and minor precipitation of opal, chalcedony, uranium minerals, gypsum, analcite and pyrite, and the formation of clastic dikes which may have been giant desiccation cracks from a post Oligocene desert Playa (Retallack, 1983). Paleosols forming at major unconformities include thick, well-differentiated layers of rock enriched in silica (Retallack, 2001).

Impact Events

Major impact events helped to shape the landscape of the Eocene-Oligocene, with late Eocene stratigraphy containing at least two iridium anomalies and deposits marked by microspherules and microtektites, shocked quartz, and coesite (Bottomley et al., 1997). Based on rock chemistry, stratigraphy, and dating, the two most identifiable and important impacts are the Chesapeake Bay impact in North America and the Popigai impact in Siberia (Cavelier et al., 1981; Stoffer and Claeys, 1987; Bottomley et al., 1997; Ivany et al., 2003).

Both impacts occurred approximately 35.5 Million years ago and happened within a few hundred thousand years of each other or less (Stoffer and Claeys, 1987; Bottomley et al., 1997). The appearance of a tektite strewn field in North America, the Caribbean

and the central Pacific to Indian Ocean is correlated within tens of thousands of years of a radiolarian extinction event at 34 Ma (O'Keefe, 1980). The Chesapeake Bay crater is most likely the source of these tektites (Stoffler and Claeys, 1987; Bottomley, 1997). The Popigai impact is evidenced as a horizon in an ODP core from the Maude Rise, and is associated with an iridium peak and shocked quartz at the lower calcareous nannofossil NP19/NP20 zone in the Eocene-Oligocene stratotype section at Massignano, Italy (Bottomley et al., 1997; Vonhof et al., 2000). In terms of the terrestrial cratering rate, this represents an unusually small time between events, as the expected repeat rate of impact structures of this size is roughly tens of millions of years (Bottomley et al., 1997).

There is some debate over the origin of the tektites, though evidence suggests they are of terrestrial origin, formed by the impact of an extraterrestrial object with the Earth's surface (Stoffler and Claeys, 1987; Bottomley et al., 1997). O'Keefe (1980) suggests the tektites were instead created from an astronomical source falling to earth with the associated impact object. If this were the case, it is reasonable to assume that a ring of material could have coalesced around the earth (O'Keefe, 1980). The lifetime of the rings would have only been a few million years (O'Keefe, 1980).

CHAPTER III: CLIMATE RECORDS

Climate often is used as a catchall term to describe temperature, precipitation, and vegetation of a region. Today we can see the effect climate has on different regions of the world-most obviously through the existence of radically different types of plants and fauna in tropical and arctic regions. Subtler climate signatures are written in the mineralogy of soils and sediments, the type of sediments deposited and degree of weathering, the type and amount of dispersed pollen, and the isotopic ratios comprising certain minerals. Given that this is true of the present, it is also true of the past. Examining the elements that are affected by climate and how those elements have changed over time delivers a record of climate change over the geologic past for any region containing one or more of these elements. Typically the records are also influenced by factors other than climate, and are subject to destruction and alteration through time. Biases in selective preservation, the qualitative nature of many climate records, and the impact of more than one influencing factor make them less reliable as stand-alone climate proxies. Each record gains more significance for climatic interpretation if it is correlated with another coeval, independent record for a given region.

Fossil Record

The fossil record is comprised of both the marine and the terrestrial record, and can be further subdivided into floral and faunal types. The species that existed, first and last appearances, diversity ratios, migration patterns, and the type and degree of preservation comprise the fossil record and serve as a snapshot of the community at any given time. This snapshot can have a profound impact on our understanding of the climate during the time of deposition and preservation. Some findings in the fossil record lend themselves to obvious climate interpretations, such as the abundance of reptiles indicating that climate must have been warm in the region. However, there are frequent disruptive forces acting on the fossil record, which complicates the interpretive process.

Marine fossil records include micro and macro fauna and flora, as well as pollen from both marine and terrestrial sources. The marine fossil record can be difficult to understand on shelf settings, where facies changes and regional tectonism often obscure evolutionary turnovers (Ivany et al., 2003). Indeed, faunal turnover patterns in neritic sections are generally influenced by regional processes such as sedimentation rates, local tectonics, and relative sea-level fluctuation (Fluegeman, 2003). Despite these difficulties, assemblages indicate what the water temperature, salinity, oxygen content, and turbidity were at the time of preservation. Correlation of isochronous records at many different sites around the world can be used to piece together the global picture, and may indicate a global climate event. On a smaller scale, migration trends, and the position of currents and isotherms can be identified, which are all indicative of changes in climate patterns (Cavelier et al., 1981).

Similar to the marine record, interpretation of the terrestrial fossil record is also fraught with difficulties. Moisture, pH, scavenging, mechanical destruction, and exposure, among other things, can all lead to destruction rather than preservation of continental fossils. Soil formation is essentially a form of early sediment diagenesis; fossil soils can therefore be a guide to the degree, nature, and location of preservational biases of animal and plant fossils (Retallack, 1983).

Bones, teeth, phosphatic coprolites, land snails, and calcareous phytoliths are all significantly more abundant in calcareous and pH neutral soils (Retallack, 1998). They also favor oxidized and moderate, rather than weakly, developed soils. This is less obvious because, though they are more often found in moderately developed paleosols, they are better preserved in weakly developed paleosols (Retallack, 1998). Fossil plants and charcoal are favored by reducing environments (Retallack, 1998). Plants are preferred in acidic and weakly developed paleosols, while charcoal is preferred in alkaline and moderately developed soils (Retallack, 1998).

From the preceding information, it is obvious that variables other than temperature are far more important in affecting terrestrial deposits than marine ones (Savin, 1980). Indeed, aridity, seasonality, and elevation strongly influence the type of vegetative biome that will develop. Differences between tropical forest, woodland, and savannah biomes can be stark or subtle in modern times, making climatic determinations based on fossil macroflora specimens challenging. By definition, tropical forests are composed of trees taller than 20m with overlapping crowns, woodlands have smaller trees and crowns that may touch but overlap little if at all, and savannahs have the most

open structure, dominated by grasses with minor to substantial woody components (Jacobs, 2004). At best, these distinctions are difficult to distinguish in the fossil record.

Within each biome are communities that vary at smaller scales, reflecting soil conditions or physical features affecting local climate (Jacobs, 2004). Rainforests receive 2000 mm of rainfall per year, have minor or no dry periods, and are evergreen (Jacobs, 2004). As precipitation decreases and dry seasons extend up to five months, the forest types change to semi-evergreen, dry forests and deciduous forests (Jacobs, 2004). Montane forests are characterized by a larger range in diurnal temperature and lower mean temperature than lowland forests, affecting species composition and size (Jacobs, 2004). Woodlands can be dense or open, depending on annual rainfall and dry season intensity, but have a grass understory (Jacobs, 2004). Savannah's may have very woody components depending the frequency of fires, which suppress the development of woody vegetation, and by grazing which promotes grass productivity (Jacobs, 2004).

Organic matter and pollen, collectively termed palynomorphs, are often found in conjunction with macroflora in the fossil record, or can be analyzed on their own. Pollen and spores range in size between 40 and 150 micrometers and tend to be concentrated in silts and clays (Eshet et al., 1988). Sediments can contain organic matter of various types, including vascular and cuticular tissue, unicellular freshwater algae, and fungal spores; marine sediments may also include dinocysts and acritarchs of marine organic-walled microplankton (Eshet et al., 1988).

Since palynomorphs are a direct result of the vegetation that existed, they are therefore a direct indicator of the climate in existence at the time the vegetation was alive. Pollen grains can often be identified to species, allowing for precise environmental

interpretations. In order to obtain the best possible date-range for deposition, care must be taken to assess whether the sediments containing the palynomorphs have been re-worked. Accurate dating is critical for tying the climate to global and/or regional events. Pollen grains, spores, and seeds originate in land vegetation and are dispersed in the land surrounding them as part of their normal method for reproduction. These can then be dispersed further by wind and fluvial processes, which can result in transport over large distances. This makes using pollen stratigraphy difficult for paleoclimate reconstruction of a specific area. Since pollen can travel long distances, it can only be used to assess larger patterns of regional trends.

When assessing the faunal fossil record, it is often tempting to link evolutionary trends to climatic events. This practice can be confounding. A growing consensus and body of evidence counters the idea that all organisms are highly sensitive to climatic changes and respond by adaptation (Prothero, 1999). While some groups, such as microplankton and land plants, respond to climate with well established behavior, other groups, particularly marine invertebrates and terrestrial vertebrates have less obvious responses (Prothero, 1999). Studies of Pleistocene mammals show that most persist through many climatic cycles of glacial/interglacial periods (Prothero, 1999). They usually respond, not by adapting or evolving, but by migrating; few evolutionary changes in land mammals can be directly attributed to climatic change (Prothero, 1999). If one can document actual migration patterns north or south to follow latitudinal climatic gradients, then an argument can be made for climate change based on the fossil record. Examining changes in fossil mammal assemblages at one site can rarely lead to a climatic interpretation.

Sediments and Paleosols

Sedimentary stratigraphy is often used as a means of interpreting the depositional environment, including the climate that the sediment formed in. The type of sediment deposited, carbonate/silicate ratios, grain size, mineral composition, hiatuses, degree of weathering, and other descriptive characteristics are commonly analyzed. Deep sea sediments often provide the most readily interpretable climate information (Savin, 1980). A hiatus in sediment deposition is the result of non-accumulation. This is either due to increased weathering, or decreased deposition. The lack of carbonate deposits usually indicates a regressive phase, whereas the sudden lack of silicates indicates transgression in a shallow marine environment. Deep marine carbonate reductions are due to deepening of the Carbonate Compensation Depth (CCD), which is dependent on seawater temperatures and pH, or an influx of terrigenous material.

Sediment stratigraphy often becomes much more complicated than this in mid-continental sequences, due to the influence of ice, fluvial and eolian processes, as well as sudden interruptions from volcanic activities. Many of the mid-continent basins are stacked with alluvial suites as nearby mountains erode. This rarely results in the “layer cake” formation that is found in deep marine sequences. “Understanding the ancient soil/landscape relationships is critical for correctly resolving the complicated histories of alluvial suites” (Kraus, 1997).

One subset of sedimentology is the study paleosols. Soil formation is a function of climate, organisms, topographic relief, parent material, vegetation, and time (McBride et al., 1968; Retallack, 1983; 1985). Soils are the zone of interaction between the atmosphere and the global crust, and therefore they may provide useful evidence for the

nature of early ecosystems on land-both complementary to the fossil record, and also in sedimentary sequences which were either too acidic or too oxidized to allow preservation of fossil plants and animals (Retallack, 1985; 2001). Using independent evidence of environments from fossil soils allows the degree to which an organism is adapted to its environment to be determined instead of assuming they are optimally adapted; therefore fossil soils provide some insight into the degree, nature and location of the biases in mammalian fossils (Retallack, 1983).

The composition, degree of development, structure, presence of burrows, root traces, color, and sedimentary relics are all studied in a fossil soil in order to ascertain the vegetation, depositional environment, and associated climate during soil formation. The presence or absence of these variables, as well as their ratio of abundance within any one paleosol allows for comparison to modern soils. However, there are distinct differences between modern and fossil soils which complicate comparison. Modern soils are defined based on the nature of their soil horizons, while fossil soils may be more hematitic and red (Walker, 1967), less rich in organic carbon (Stevenson, 1969), compacted, eroded or metamorphosed. This can make the horizons difficult or impossible to distinguish. Identifying a fossil soil from a modern soil classification system inherently implies the same environment of formation, so care must be taken when applying this method.

Differentiating paleosols from other sedimentary sequences requires in depth analysis. However, certain features are fairly indicative of soils, and can be very useful in distinguishing paleosols in the field. Those are: usual and gradational contacts from one horizon to the next below a sharply truncated upper fossil soil surface, and a massive and bioturbated appearance (Retallack, 1983). Modern soil textures are also often visible in

paleosols. These include pisolites, lumps, clay skins on detrital grains, poor sorting, randomly oriented clasts, irregular vertical tubular structures (fossil root traces and burrows), complex cracking and veining (peds and cutans), distinctive petrographic textures (sepic plasmic fabrics), and relic sedimentary features (grain size changes) (McBride et al., 1968; Retallack, 1983; 1985).

Chemical analysis of the suspected paleosol can further support the classification. A common characteristic of soils is surficial depletion of the plant nutrients Potassium oxide (K_2O) and Phosphorous pentoxide (P_2O_5) in proportion to their degree of development (Retallack, 1983). Modern soils also typically have a high and sharp 15\AA smectite peak in their upper horizons (Retallack, 1983). These features were used to identify paleosols in the Badlands National Park, as previously discussed.

Paleosols are formed through cumulative or non-cumulative processes. The formation of cumulative paleosols occurs when pedogenesis keeps pace with sediment accumulation, and deposition is slow and gradual (Kraus, 1997). In contrast, non-cumulative soils develop in succeeding stages, where each stage gradually replaces the previous stage-e.g. the old A horizon becomes the B horizon, and the top of the profile becomes the new A horizon (Nikiforoff, 1949). During development, the soil develops certain characteristics that can not be destroyed during the following stages (Nikiforoff, 1949). Therefore, the evolution, degree, and nature of paleosol development can be assessed by analyzing the soil characteristics found in the parent material, which is usually mildly preserved in the C-horizon (Retallack, 1983). Comparison to modern soil can yield a quantitative estimate of the minimum time for development (Birkeland,

1974), while comparison of the degree of expression of soil features to sedimentary features can yield the relative time for development of a paleosol (Retallack, 1983).

Certain formations within the paleosol can indicate periods of deposition or non-deposition. Concentrations of burrow-like irregularities on the tops of many paleosols and elongate nodules (burrow fills) constitute trace fossil evidence for periods of non-deposition and soil formation (Retallack, 1983). Weakly developed paleosols indicate rapid sedimentation (Kraus, 1997). The rate of sediment accumulation can be calculated from the minimum number of paleosols in a sequence and the thickness of a sequence assuming: 1) no or few prior paleosols have been overprinted by the visible soils, 2) there has been little erosion and destruction of paleosols, 3) once a soil reached a given degree of development it was covered by more sediment, and 4) long diastems are located between, rather than within, sequences used for calculations (Retallack, 1983).

Further analysis can lead to an understanding of environmental conditions during formation. Moisture and temperature are the two most important aspects of climate in controlling soil properties-moisture because water is involved in most of the physical, chemical, and biochemical processes that occur in a soil and affects the weathering and leaching conditions; and temperature because it influences the rate of chemical and biochemical processes (Birkeland, 1974). The climate-soil relationship, or climofunction, is best demonstrated through a comparison of the U.S. Soil map with a map of precipitation and temperature, as most soil orders and suborders are restricted to certain climatic regions (Birkeland, 1974).

Given that paleosol formation is essentially a form of early sedimentary diagenesis, a clear distinction between early and late diagenesis is important for a

paleoenvironmental study of fossil soils (Retallack, 1983). Even when diagenesis is accounted for, it is difficult to detect sudden or cyclical climatic changes due to the gradational changes in climatically sensitive features from one paleosol to another, gaps in time between paleosols, and local paleoenvironmental effects (Retallack, 1983). Time-averaged trends in paleoclimate can be assessed, though, primarily through examining the mineralogical composition.

The main soil properties that correlate with climate are organic-matter content, clay content, kind of clay minerals, color, the presence or absence of CaCO_3 , and depth to the top of the CaCO_3 bearing horizon (Birkeland, 1974).

Oxidized soils are reddish in color, and are composed of minerals such as ferrihydrite, goethite, hematite, hornblende and biotite (Retallack, 1983; 1985; Walker, 1967). Reduced fossil soils are drab colored (bluish or greenish gray), may have the purple tinge of ferruginized horizons, gley mineralization such as pyrite or siderite, lack weathering features, and may be carbonaceous (Birkeland, 1974; Retallack, 1983; 1985). Leached A2 horizons may also be drab or light grey, but in formerly oxidized soils, they lack carbonaceous minerals (Retallack, 1983; 1985; Walker, 1967). A paleosol may also be drab colored if it underlies a layer that underwent reduction. For example, the Devonian paleochannel deposits in New York contain pyritized logs, which transmitted reducing conditions to the underlying red claystone (Retallack, 1985).

Soils with poor drainage, high water tables or leaching can contain a high number of salts, and may have formed in a similar manner to soils with calcium carbonate (CaCO_3) (Birkeland, 1974). If conditions are part oxidizing and part reducing, likely due to fluctuating water table levels or water content, a mottled color scheme of gleyed, with

some brown, yellow and red compounds will occur (Birkeland, 1974). Low pH soils are non-calcareous; as they become more alkaline they also become progressively more calcareous (Birkeland, 1974; Retallack, 1983). Soils with a pH of 9 or 10 may have zeolites, gypsum and other evaporate minerals (Retallack, 1983).

The most obvious diagnostic feature of a soil in terms of the vegetation it supported is the distribution of organic matter with depth, which varies based on the plants growing (Birkeland, 1974). The use of organic content for re-constructing paleoclimate is limited, due to its ability to reach steady state with current and changing conditions quite quickly (Birkeland, 1974).

Color can be quite deceiving when interpreted on its own as indicative of a particular climate. Iron can be redistributed by plant toxins, dehydrated during early burial, or leached out of A and B horizons into the lower C horizons, or even from one paleosol or sediment to an underlying deposit (Walker, 1967; Retallack, 1985). It is likely that original oxidation and formation of ferric oxyhydrate minerals occur during formation, but reddening is the result of late diagenetic altering. Staining is caused by mobilization of iron through filtrating water; wherever this water is in contact with iron silicates, alteration of the iron-bearing grains may occur as long as the hydraulic gradient allows removal of soluble products (Walker, 1967). Mobilized iron will remain in solution until encountering sediments with enough oxidizing material, whereupon $\text{Fe}(\text{OH})_3$ will precipitate and ultimately form hematite, turning the associated beds red (Walker, 1967). Note that climate is not an influencing factor in the mobilization and precipitation of iron. In situ alteration is notable in the field when older underlying

sediments are more strongly pigmented than the younger overlying sediments, which also tend to be rich in iron bearing silicates (Walker, 1967).

A useful climofunction for estimating mean annual rainfall from paleosols is the depth to the top of the calcareous nodular horizon. Depth to calcareous nodules is not the same as the depth to leaching of carbonate in non-calcareous soils of humid climates, depth to calcareous stringers or horizons of very old soils, nor with depth to pore filling cements; calcite nodules are robust (Retallack, 1992). The primary obstacles with this climofunction are compaction during burial and pre-burial erosion (Retallack, 1992). Corrections for compaction can be made using geological information on depth of burial and standard compaction curves while surface erosion can be assessed based on preservation of root traces and soil structures (Retallack, 1992). Though the climofunction varies in surface soils depending on a variety of conditions, deeper horizons are consistently found in wetter regimes among soils of different climate, parent material and time for formation (Retallack, 1992).

From these compositional analyses, trends in soil formation with regard to the climate under which they formed can be found. Tropical soils are generally more red and clayey, less rich in organic matter, and more deeply and thoroughly weathered than soils of temperate climate (Birkeland, 1974; Retallack, 1985). As discussed previously some of these features, such as red staining, are attributable to multiple factors. Increased clay depth and weathering are also related to the time available for soil formation, mean annual rainfall, and temperature, while decreased organic matter may be the result of burial (Retallack, 1992). Tropical forest regions are also characterized by iron and aluminum oxides and hydroxides predominantly close to the surface (Birkeland, 1974),

and they may contain calcite concretions and stringers if there is a pronounced dry season (Mcbride et al., 1968).

Soils that are drab in the upper part but stained red in the lower part of the profile from modern oxidation of original pyrite have been found forming today under coastal salt marsh and mangrove vegetation with similarly gleyed sulphidic soils (Retallack, 1985). Humid climates enable the development of soils which contain a band of humus (Stevenson, 1969). In humid forests, the humus accumulation zone is found in the B-Horizon, while in humid grasslands the humus layer is formed at depth in the soil profile (Stevenson, 1969). In any soil, the humus is the most susceptible to change, thus much of the humus has been lost in paleosols (Stevenson, 1969). Bioturbation of the surface of a paleosol also indicates a humid climate (Retallack, 1985). The presence of calcite, especially at depth, indicates a humid climate (Retallack, 1985).

Soils in semiarid and arid regions commonly have CaCO_3 and K horizons (Birkeland, 1974). Red beds with immature clay, high pH, and carbonates throughout the profile indicate formation in a hot, dry climate (Walker, 1967), though thick hard caliche crusts are known to form also in sub-tropical soil subject to alternate wetting and drying—as preciously mentioned (Mcbride et al., 1968). Paleosols in hot dry climates are particularly susceptible to red bed formation due to commonly low water tables, sparse vegetation, and alkaline ground water (Walker, 1967). Low precipitation can be inferred by the presence of associated weathering products of rain, or lack thereof. Where leaching is slower, precipitation rates are lower and the composition of the resulting sediment is immature clays, incomplete alteration of iron silicate minerals such as hornblende, and calcium carbonate—particularly in the upper layers (Walker, 1967).

The tundra, where permafrost exists with low precipitation and low temperatures, creates an A horizon with a fairly high organic content overlying gleyed horizons (Birkeland, 1974). The polar desert soils are quite similar to hot desert soils, with little vegetation providing organic content, and low moisture leading to slow rates of weathering and increased salt content and alkalinity (Birkeland, 1974).

A steppe is marked by increased precipitation and decreased evapotranspiration, creating thick A horizons with vegetation rich in organic matter and slight leaching; only the most arid regions have CaCO_3 (Birkeland, 1974). Teepee structures and surface cracking are indications of variations in rainfall, indicating seasonality (Retallack, 1985). Altering layers of ferruginization and calcification also indicate seasonality, with ferruginization occurring during the rainy season and calcification during the dry season (Retallack, 1985). Seasonal water-logging indicates formation on low levees and swales flanking a coastal stream or deltaic distributary (Retallack, 1985).

Using locations where climate was constant, the changes in vegetation can be assessed against paleosol development, and thus conclusions regarding vegetation's impact on soils can be assessed. Thicker profiles that are well differentiated, particularly with large root traces, indicate forest coverage (Retallack, 1985). Forest A horizons are thin, with organic matter being sparse at depth, whereas grassland A horizons are thicker with organic matter throughout (Birkeland, 1974).

Surface layers in grasslands have higher clay content than in forests (Birkeland, 1974). Thinner, calcareous soils with smaller root traces indicate herbaceous or bushy vegetation (Retallack, 1985). Savanna vegetation is indicated by yellow to brown, moderately to weakly developed, granular cracking and fine, branched root traces

(Walker, 1967; Retallack, 1985). Some plants secrete opal, which conforms to the shape of the cell from which it formed and is known as a phytolith (Birkeland, 1974).

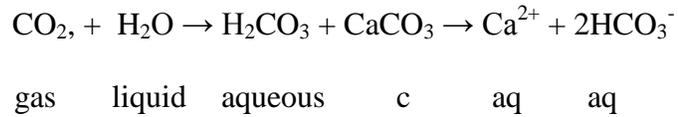
Grasslands produce more phytoliths per area than forests (Birkeland, 1974). Given all of the above mentioned difficulties in utilizing any one line of evidence to place a paleosol in a particular climatic environment of formation, a broad based view can help in reaching generalized conclusions. By taking into account the entire composition and relationship of different factors to each other within the same soil and surrounding soils, interpretations regarding climate can be made.

Soil CaCO_3 has been mentioned several times thus far. It is found in nearly every environment of deposition, and has been used as an indicator for relative rates of precipitation. Soil calcite is worth discussing in depth due to its robust nature and usefulness in unlocking climatic signals. CaCO_3 is found in many different environments because wherever calcium, carbonate and water exist, only favorable conditions are needed for precipitation of the mineral. Calcium and calcium carbonate are typically derived as weathering products of the nearby deposits, but can be transported from external sources. Air-fall is an important factor for distribution of Ca^{2+} for many arid localities (Birkeland, 1974).

As previously discussed, soils with a low pH are non-calcareous; as they become more alkaline they also become progressively more strongly calcareous (Birkeland, 1974; Retallack, 1983). The presence or absence of calcite can therefore be very helpful in determining the relic pH, saturation state, and degree of drainage for a paleosol.

The origin of a calcic horizon involves the equilibrium of carbonate-bicarbonate; increases in soil gas carbon dioxide (CO_2) or decreases in pH will drive the below

equation to the right-dissolving bicarbonate and mobilizing calcium (Ca) and bicarbonate (HCO₃) as water percolates through the soil (Birkeland, 1974).



The process of calcium carbonate dissolution and precipitation begins in the surface layers of the soil. Root and organism activity occurs in the upper A horizon, along with increased water percolation and organic matter decomposition (Birkeland, 1974). This increased activity results in soil pore CO₂ partial pressures that are 10 to 100 times higher than in the atmosphere (Birkeland, 1974). Increased CO₂ partial pressures decrease pH, and upon contacting water it turns to carbonic acid (H₂CO₃) (Birkeland, 1974). In addition to increased CO₂ partial pressures in the upper soil horizon, mineral weathering is also taking place, resulting in the liberation of calcium. CaCO₃ is preferentially dissolved due to the presence of H₂CO₃ as mentioned. Water percolating through the soil mobilizes these two compounds until they are at lower levels of CO₂ partial pressure-typically below the biotic zone (Birkeland, 1974).

As the water percolates deeper into the soil horizon, ion concentrations increase and CO₂ partial pressure decreases. CaCO₃ is precipitated once a favorable balance is struck (Birkeland, 1974). The position of the CaCO₃ horizon is therefore related to depth of leaching, because CO₂ is less soluble in warm water than in cold water, CaCO₃ solubility decreases with rising temperatures, directly linking the depth of calcic horizons to climate (Birkeland, 1974). Dissolution occurs when the volume of water moving

through the soil increases, as long as the water is not already saturated with CaCO_3 (Birkeland, 1974). Decreasing CO_2 or increasing pH leads to precipitation of carbonate, as does an increase in ion content such that saturation is reached (Birkeland, 1974).

Though depth to calcite is often used as a proxy for precipitation, leaching of carbonate depends on the depth of wetting and the production of carbonic and other acids by soil biota, which can be quite different in the geological past compared with modern soils (Retallack, 1985). Additional lines of evidence such, as abundance of burrows and weathering of iron-bearing minerals or other climate records should be examined before precipitation estimates are made.

As previously discussed, fossil soil features can and have been used to interpret past vegetation and climatic changes for specific regions. However, these attempts are marred with problems that can lead to inaccurate conclusions and at best give a qualitative estimate. Soil features can best be used to assess changes in climate and vegetation when studied over a transect that underwent dramatic change. A transect that experienced a more subtle or gradual change may not record any of those changes in its soil features. Furthermore, changes from arid to wet conditions may not be recorded at all due to dissolution and weathering. Problems with weathering, time of formation, time to reach steady state, etc. all come into play and reduce certainty and accuracy in using paleosol features to interpret climatic changes. More quantitative methods are necessary to support any qualitative findings.

Isotope Record

The most established and reliable proxy record used to define the timing and magnitude of global climate change is the oxygen isotopic ratio ($\delta^{18}\text{O}$) of benthic foraminiferal calcite obtained from deep sea cores (Lear et al., 2000). Biogenic precipitation of carbonate follows essentially the same chemical process as the formation of pedogenic carbonate, and can be described by the below equation, provided by Owen and Rea (1985):



Biogenic precipitation of CaCO_3 is the only significant process that balances Ca^{2+} inputs to the ocean, and results in the formation of CO_2 (Owen and Rea, 1985). It is notable that enrichment of seawater with Ca^{2+} occurs during increased sea-floor ventilation, as sea-water reacts with hot basalt and is concomitantly depleted in Mg^{2+} (Owen and Rea, 1985). The relationship of increased Ca^{2+} leading to greater biogenic precipitation, which in turn releases CO_2 , further links tectonic activity to climate change, as will be discussed later (Owen and Rea, 1985).

Given that carbonate is partially composed of oxygen, the oxygen isotopic record is readily available in any carbonate deposit. $\delta^{18}\text{O}$ (spoken as “del-18-O”) is the deviation of the oxygen isotope ratio R of a sample from that for a standard, typically Pee Dee Belemnite (V-PDB). This is calculated as:

$$\delta^{18}\text{O} = 1000 \times (\text{R}_{\text{sample}} - \text{R}_{\text{standard}}) / \text{R}_{\text{standard}},$$

$$\text{where } R = {}^{18}\text{O}/{}^{16}\text{O}$$

Carbonate forming animals take up the oxygen isotopes from the water in which they live when making the carbonate that forms their tests. These tests can then be analyzed for their abundance of each oxygen isotope in order to extrapolate the temperature of the ocean at the time of formation. The fundamental premise of the analysis is that ${}^{16}\text{O}$, which is lighter than ${}^{18}\text{O}$, is more easily evaporated from oceanic waters, enriching oceans in ${}^{18}\text{O}$ relative to meteoric water. ${}^{18}\text{O}$ is thus expected to be in abundance in marine organismal carbonate tests-more so as temperatures cool, because a warmer atmosphere increases the rate of evaporation, thus taking up ${}^{16}\text{O}$ and ${}^{18}\text{O}$. Furthermore, as climate cools and glaciers form, ${}^{16}\text{O}$ is preferentially removed from the system as it accumulates in glacial ice (Cavelier et al., 1981). Even meteoric waters, and the carbonate they precipitate, will become enriched in ${}^{18}\text{O}$.

When the isotopic ratio is displayed on a graph against time, major shifts and minor perturbations can be found across the geologic record. $\delta^{18}\text{O}$ enrichments are noted as shifts towards more positive values, while $\delta^{18}\text{O}$ depletions are noted as shifts towards more negative values. Recognition of these shifts can help constrain the age of particular deposits and correlate disparate sections; the $\delta^{18}\text{O}$ curve may be the ideal tool for recognizing global isochronous time planes (Ivany et al., 2003)

Local or regional changes in upwelling, circulation, precipitation, seasonality or temperature can affect the $\delta^{18}\text{O}$ of terrestrial and shallow marine carbonates. The $\delta^{18}\text{O}$ of benthic forams is powerful because the deep ocean is insulated from large seasonal,

latitudinal, and geographical variations in temperature and salinity, making it more representative of global change than surface ocean or continental records (Lear et al., 2000). The relationship between $\delta^{18}\text{O}$ and global temperature and ice sheet growth has been established. Generally $\delta^{18}\text{O}$ increases as temperature decreases at a rate of 0.25‰/°C; and as the mass of continental ice increases at a rate of 0.1‰/10 meters of sea-level change (Zachos et al., 2001). Shallow water organism and continental deposits are analyzed in terms of their oxygen isotopes to help constrain local climate changes, and also to aid in untangling some of the issues regarding the temperature record, as will be discussed later.

The two main agitators to the analysis are the role that global ice volume plays in fractionation, as well as metabolic fractionation by the organisms themselves. This is because the $\delta^{18}\text{O}$ of marine carbonates are dependent on temperature and the $\delta^{18}\text{O}$ composition of seawater ($\delta^{18}\text{O}_{\text{sw}}$) (Lear et al., 2000). The composition of sea-water is affected by the global ice volume as well as temperature, as previously mentioned. This results in an increase in $\delta^{18}\text{O}$ in the oceans, which is not strictly related to temperature (Cavelier et al., 1981). Especially at high altitudes, continental glaciers can form even when there is no global decrease in temperature-due to the presence of microclimates. Therefore, a $\delta^{18}\text{O}$ signal recorded in marine carbonate reflects both temperature and ice volume. It is important that we sort out glacial volume and deep-sea temperatures over earth's history, since they each have far reaching impacts (Dwyer, 2000). “If an independent record of deep-sea temperature was available, this complexity in the benthic foraminiferal calcite $\delta^{18}\text{O}$ record could be an important advantage in determining global ice volumes” (Lear et al., 2000).

Many indirect methods attempt to extract the global ice budget from deep-sea $\delta^{18}\text{O}$, and have resulted in very different estimates of the onset of continental-ice sheet accumulation (Lear et al., 2000). Evidence supporting major episodes of glaciation includes ice rafted debris, weathered clay mineral compositions, microfossil assemblages, sea level changes and hiatuses (Miller, 1992; Lear et al., 2000). However, these records can not estimate ice volume or determine the cause of the increase or decrease in ice volume. These methods are also often incomplete and ambiguous (Lear et al., 2000).

The partition coefficient of Mg^{+2} into inorganic calcite correlates strongly with temperature, allowing for the use of the Mg/Ca ratio to establish paleothermometry (Lear et al., 2000). The use of the Mg/Ca temperature record to calculate $\delta^{18}\text{O}_{\text{sw}}$ component of $\delta^{18}\text{O}$ has been successfully applied by Dwyer et al., 1995 to ostracode records of late Pliocene and Quaternary age; by Hastings et al., 1998 to planktonic forams of the Quaternary for sea-surface temperatures; to benthic Quaternary forams for deep-sea temperatures by Rosenthal et al., 1997; and to Eocene-Oligocene benthic forams (Lear et al. 2000). This independent assessment of $\delta^{18}\text{O}_{\text{sw}}$ allows for the temperature and ice-volume effects to be untangled from the global $\delta^{18}\text{O}$ curve.

Simple geochemical models, together with known fluxes of Mg and Ca into and out of the oceans, are used to estimate past seawater Mg/Ca characteristics (Lear et al., 2000). Rivers transfer Mg and Ca from the continent to the ocean. Submarine hydrothermal processes remove some of the Mg by exchange for Ca from the oceanic crust. Therefore, Ca from rivers and hydrothermal sources is removed only during carbonate deposition; whereas Mg, derived only from rivers, is removed during dolomite production and hydrothermal activity (Lear et al., 2000). To translate Mg/Ca records into

temperatures, the modern Mg/Ca temperature calibration from *C. floridanus* and a species adjustment to the reference species *O. umbonatus* is used (Lear et al., 2000).

The application of Mg/Ca paleothermometry to longer time scales introduces problems: diagenesis effects, changes in seawater Mg/Ca, and uncertainties in applications within Quaternary paleoceanography (Dwyer, 2000; Lear et al., 2000). It is important to note, the Mg/Ca relation to temperature may not be as foolproof as it is presented here—factors other than temperature may affect the amount of Mg taken up during shell formation (Dwyer, 2000). These include enigmatic ontogenetic and interspecies variations, and preferential dissolution and re-crystallization of Mg during diagenesis (Dwyer, 2000).

Carbonate tests are also composed of carbon isotopes. The $\delta^{13}\text{C}$ (spoken as “del-13-C”) ratio can be used to interpret paleoclimate. While benthic $\delta^{18}\text{O}$ is used as a proxy for temperature and ice volume variations, benthic $\delta^{13}\text{C}$ is used as a proxy for water chemistry such as ventilation changes, the ratio of deposition of organic carbon to carbonate in the oceans, and long-term changes in the relative size of carbon reservoirs in the biosphere—including relative abundance of atmospheric CO_2 , nutrient influx from terrestrial sources, and sea-level changes (Cavelier et al., 1981; Diester-Haass and Zahn, 1996; Hurley and Fluegeman, 2003; Zanazzi et al., 2007). Like $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ is the deviation in permil of the carbon isotope ratio of a sample from that of a standard (V-PDB):

$$\delta^{13}\text{C} = 1000 \times (\text{R}_{\text{sample}} - \text{R}_{\text{standard}}) / \text{R}_{\text{standard}},$$

$$\text{where } \text{R} = {}^{13}\text{C} / {}^{12}\text{C}$$

$\delta^{13}\text{C}$ variations in carbonates can be correlated with global sea level changes—a negative ratio is equivalent to a regressive phase while a positive ratio indicates a transgressive phase (Cavelier et al., 1981); $\delta^{13}\text{C}$ is also a useful indicator of deep-sea venting. Typical deep-sea benthic $\delta^{13}\text{C}$ values range from approximately -0.5‰ to 1‰ (Hurley and Fluegeman, 2003).

Epicontinental marine fauna may be used for stable isotope analysis because they will react very quickly to climatic variations (Cavelier et al., 1981). However, epicontinental marine fauna are subject to interferences of dilution, desalination, and even confinement (Cavelier et al., 1981). Oxygen and carbon isotope curves have also been derived from sources including stalagmites (Denniston et al., 2007); fossil teeth (Zanazzi et al., 2007) and paleosol carbonate nodules (Cerling, 1991; Bowen et al., 2001; Mullin and Fluegeman, 2009).

The isotopic chemistry of authigenic soil minerals can be a sensitive record of climatic and environmental conditions; pedogenic carbonate from paleosols is a valuable and widely used source of paleoclimatic and paleoenvironmental proxy information (Bowen et al., 2001). The oxygen isotopic composition of modern soil carbonate is well correlated with the isotopic composition of local meteoric water (Cerling, 1984). $\delta^{18}\text{O}$ is typically interpreted as recording changes in paleoclimatic variables, while $\delta^{13}\text{C}$ responds to more general paleoenvironmental shifts (Bowen et al., 2001). Temporal variation in the $\delta^{18}\text{O}$ value of precipitation has commonly been interpreted to reflect variation in mean annual temperature, but this is not applicable in all environments. Where annual rainfall is less than 35 cm per year, the oxygen isotope composition of pedogenic

carbonates can be isotopically enriched due to significant evaporation and diffusional mixing with the atmosphere as soils have very low respiration rates (Cerling et al., 1989).

Screening for kinetic effects, alteration and pre-infiltration evaporative enrichment of ^{18}O in meteoric waters is accomplished by the existence of identical values and trends in multiple coeval samples (Denniston, et al., 2007). Unless pedogenic carbonates undergo dissolution and re-crystallization, isotopes of the infiltrating water should reflect the oxygen isotopic composition of infiltrating water and the temperature at which the calcite precipitates from the water (Denniston et al., 2007).

Carbon in soil carbonate is derived from soil CO_2 , which is a mixture of carbon dioxide from the atmosphere and organic decomposition, and soil-respired CO_2 , derived from root respiration (Bowen et al., 2001). The combination of soil respiration, microbial activity, and influence from atmospheric CO_2 , leads to heavier isotopic distributions in soil CO_2 , compared to atmospheric CO_2 , (Cerling, 1991). Many variables determine the $\delta^{13}\text{C}$ value for soil carbonate, but the most important are depth in the soil profile, the fraction of C_4 biomass, and $\text{P}(\text{CO}_2)$ of the atmosphere (Cerling, 1991). Depth in the soil profile is significant only in the upper 20cm in most soils; below this the carbon isotopic composition is essentially constant (Cerling, 1991). Additionally, in more recent deposits (~8 Ma), $\delta^{13}\text{C}$ can show the variation in plant land cover from C_3 to C_4 plants (Prothero, 1999). C_4 plants are grasslands which typically dominate most modern temperate and tropical latitudes (Prothero, 1999). C_3 plants have a $\delta^{13}\text{C}$ range of -7‰ to -14‰, while C_4 plants have values higher than -7‰ (Prothero, 1999). The existence of C_4 plants older than Miocene is not supported; therefore, in the absence of C_4 plants, the isotopic

composition of soil carbonate is a potential indicator of atmospheric $P(\text{CO}_2)$ levels (Cerling, 1991).

Soil CO_2 , refers to the concentration of CO_2 , in a particular soil volume and has units of ppmV; soil-respired CO_2 , represents a flux and is the amount of CO_2 , passing through a horizontal plane in the soil, such as the soil-air interface, and has units of $\text{mmoles/m}^2/\text{hr}$ (Cerling, 1991). Soil CO_2 and soil-respired CO_2 have very different isotopic compositions. The concentration of soil CO_2 is controlled by the production rate of CO_2 (which is a function of depth in the soil) and diffusion through the soil to its upper boundary (the atmosphere) (Cerling, 1991). The diffusion coefficients of $^{12}\text{CO}_2$ and $^{13}\text{CO}_2$ differ by 1.004, resulting in enrichment in $^{13}\text{CO}_2$ in soil CO_2 compared to soil-respired CO_2 ; the isotopic composition of soil CO_2 is 4.4‰ or more enriched in $^{13}\text{CO}_2$ relative to soil-respired CO_2 at all levels in the soil (Cerling, 1991).

The isotopic composition of soil carbonate tracks the isotopic composition of soil CO_2 , (Cerling, 1991). The combined effects of diffusion of soil-respired CO_2 , and isotopic fractionation between soil CO_2 and carbonate ($\sim 10.5\%$), predicts a 14‰ to 17‰ difference between soil-respired CO_2 and soil carbonate precipitate in isotopic equilibrium (Cerling, 1991). This effectively means that carbonate $\delta^{13}\text{C}$ will be roughly 15‰ enriched in $^{13}\text{CO}_2$ compared to that of the overlying vegetation. Therefore, Eocene soil respired CO_2 overlain by C_3 vegetation with a $\delta^{13}\text{C}$ range of -7% to -14% would be expected to have a range of -22% to -29% . Fractionation and diffusion effects would essentially nullify this as carbonate precipitates, and $\delta^{13}\text{C}$ values are expected to be within the range of -5% to -15% .

Vegetation $\delta^{13}\text{C}$ should track the $\delta^{13}\text{C}$ of atmospheric CO_2 with a consistent fractionation of approximately -19‰ for C_3 photosynthetic pathway plants (Bowen et al., 2001). There is limited variation in the isotopic composition of individual C_3 plants (Cerling, 1991). Some C_3 plants are depleted several permil in ^{13}C . This is most likely in closed canopy forests where $P(\text{CO}_2)$ is significantly higher than the ambient atmosphere, and the atmospheric CO_2 is several permil depleted in ^{13}C because the degassed soil CO_2 is not easily dissipated under closed canopy conditions (Cerling, 1991). However, soil carbonate is generally not preserved in these types of soils.

There is little evidence of isotopic kinetic effects in soil carbonate formation (Cerling, 1991). However, low soil porosity, high temperatures, and high productivity will create paleosol carbonates that tend to underestimate the atmospheric $P(\text{CO}_2)$. A conservative estimate of the range of atmospheric CO_2 in paleosols is taken from the average value for a series of soils, skewed toward the more negative values (Cerling, 1991).

CHAPTER IV: CAUSES OF CLIMATE CHANGE

It should be apparent by now that no single event is responsible for changing the climate at either a global, regional or local level. Several causes exist for global climate change at the Eocene-Oligocene: tectonics, weathering rates, albedo, and impact events. The interplay between all of these is particularly important in shaping major global events. For global changes to occur, dramatic upheavals in atmospheric chemistry and/or oceanic circulation patterns appear to be required. Important factors impacting climate change in the mid-continent include geography, atmospheric CO₂, solar radiation (Sloan and Barron, 1992), and regional air mass patterns. Each of the processes described below has some impact on the input or removal of either solar radiation, or greenhouse gases and their distribution across the oceans and landmasses.

Tectonics

Increased volcanism, changing geographic locations of continents, and uplift of land masses are the most important contributions of tectonic activity to global climate change. Generally, increased worldwide plate movements increase the amount of CO₂ supplied to the atmosphere and hydrosphere (Briggs, 1987). As continental positions change, oceanic circulation patterns are affected. Changing geography can often lead to

continental collision, creating mountains and thus exposing more continental surface to the effects of weathering. Uplifted land masses can also impact atmospheric and fluvial patterns, creating variations in local temperatures and aridity.

Increased volcanism can occur as the creation of spreading centers or increased spreading rates at mid ocean ridges (MOR), increased island building, or increased terrestrial volcanic activity. All volcanic activity releases greenhouse gasses such as methane and CO₂ into the atmosphere. Increased MOR activity and volcanic island building results in thermally buoyant oceanic crust displacing ocean water onto land, which in turn forms epicontinental seas that reduce the amount of land exposed to weathering (Briggs, 1987). The net effect of increased CO₂ supply and decreased removal is a rise of CO₂ in the atmosphere and hydrosphere (Briggs, 1987). In addition to CO₂ being released through venting, hydrothermal activity may lead to increased CO₂ production by biogenic organism (Owen and Rea, 1985). Hydrothermal venting increases the amount of Ca²⁺ available for biogenic carbonate precipitation, which in turn leads to the production of CO₂, as previously discussed. Owen and Rea estimate that seafloor hydrothermal activity accounts for ~ 29% of total ocean contribution to atmospheric CO₂ when this process is accounted for (1985).

Terrestrial volcanism is typically explosive, resulting in sudden releases of ejected ash and particulate matter into the atmosphere. Ejecta can persist in the atmosphere for years or even decades, and spread across the globe. The effect of globally spread and persistent ejecta in the atmosphere is insulation of the earth's surface from solar radiation. Particulate matter in the upper atmosphere reflects incoming radiation back into space, preventing heat from reaching the surface, and leading to cooler global conditions.

If increases in plate movements generally lead to increases in CO₂ in the atmosphere, then the reverse must also be true: a reduction in global tectonic events would reduce the input of CO₂ into the atmosphere, lower relative sea-levels, and marginalize or eliminate epicontinental seas therefore exposing more land to weathering. Increased weathering removes CO₂ from the atmosphere, lowering the net amount of CO₂ in the atmosphere-hydrosphere.

Uplifted land can also cause significant changes in atmospheric circulation patterns, which can have a dramatic impact on regional climate. Sloan and Barron (1992) conducted several atmospheric general circulation models alternating various combinations of sea-surface temperature (SST) and topography characteristics to determine effects on interior continental climate. Results revealed that an elevation change of only 500-1000 meters is enough to affect surface temperature distributions (Sloan and Barron, 1992). Substantial temperature responses to elevation changes occurred in many regions, temperatures rose 3-8K (Sloan and Barron, 1992). Altered circulation patterns caused temperature responses even in regions with no elevation modification (Sloan and Barron, 1992). Global mean surface temperature changed up to 0.2K purely through topographic modifications (Sloan and Barron, 1992).

Weathering

As previously mentioned, weathering rates play a direct role in the ability for CO₂ to be removed from the atmosphere. The weathering of rocks leads to the removal of atmospheric carbon dioxide (CO₂) (Ivany et al., 2003). Computer models by Berner et al (1983) and Lasaga et al (1985) examine the carbonate-silicate geochemical cycle and

organic carbon and sulfur reactions, respectively (Briggs, 1987). Results indicate the CO₂ in the atmosphere is highly sensitive to changes in sea-floor spreading rate and continental land area (Briggs, 1987).

Glacial formation

The volume and distribution of glacial ice can impact surface albedo (discussed below), surface air temperature gradients, atmospheric circulation patterns and global sea level (Dwyer, 2000). Given that all these proxies for climate are affected by ice sheet growth, it is not hard to understand the confusion in $\delta^{18}\text{O}$ records. Large masses of ice can cool the air flowing over them, resulting in cooler temperatures in regions not immediately adjacent to glaciers. Cooler air masses may also change overall circulation patterns by creating air currents and jet streams that affect weather patterns over continents.

If ice volume is large enough, global sea levels are lowered. This exposes more land to erosion, resulting in removal of atmospheric CO₂. Glaciers are themselves efficient instruments of erosion; even small glaciers have impressive erosional power. A positive feedback cycle exists between erosion and glaciers: as glaciers increase in number, area or volume, they increase weathering rates, which in turn cool the atmosphere and lead to the creation of more glaciers.

Albedo

Albedo is defined as the percentage of incoming radiation that is reflected by a natural surface, such as the ground, ice, snow, water, clouds, or particulates in the atmosphere (Bates and Jackson, 1984). Therefore, sea-level drop or tectonic uplift

exposing more land, as well as increased tectonics leading to more particulates in the atmosphere or increased ice production, could have significant impacts on the earth's albedo. As albedo increases, the amount of solar radiation able to either penetrate or stay trapped within the atmosphere is decreased, and the earth subsequently cools. Therefore, increasing glacial surface area creates another positive feedback cycle with respect to albedo-more surface area reflects more solar radiation, cooling the atmosphere, leading to the creation of more glacial surface area.

Oceanic and atmospheric circulation

Oceans are the world's major heat sink. Water reacts slowly to changes in ambient temperature, but ultimately will reflect dominant temperature regimes. As water cools, it sinks, and as it warms, it rises. This creates an advective motion, which is also mirrored in the atmosphere. Today's patterns show that cold air and water originate at the poles and sink. Sinking water from the poles travels along the bottom of the ocean towards the equator. Atmospheric patterns are more complicated due to inhibiting factors such as topography, microclimates, and jet streams. Tectonic uplift, creation of cold, dense water, and influxes of low salinity water into the salty oceans all have an impact on oceanic circulation.

Deep-sea temperature is fundamental in controlling physical, biological, and chemical processes in the ocean (Dwyer, 2000). Changes in deep-sea temperature lead to changes in seawater density and sea level, and reorganization of heat, salt and biogeochemical budgets through changes in density-driven circulation mechanisms (Dwyer, 2000). Cold deep-sea water also plays a role in stabilizing gas hydrates, which

can contribute to global warming events when released. Deep-sea temperatures are reflective of the sea-surface temperature at the site of deep-water formation.

The sinking and traveling of cold water along the bottom of the ocean is an important driver for the movement of warm surface water. Warm water is able to keep local environments equable, such as modern day Wales which is supplied with warmth by the Gulf Stream. If the transport system is shut down, or moved, local and even global climates can be strongly influenced. Sloan and Barron (1992) modeled the results of changing sea-surface temperatures with respect to interior continental climate. They found that as meridional SST gradients became steeper, the largest continental temperature changes occurred poleward of 60 degrees. Winter temperatures were more sensitive to change than summer temperatures, while the global surface temperature response was minimal. Large changes in SST resulted in only very small interior continental surface temperature changes.

Impact events/ extra terrestrial occurrences

Major impact events can play a significant role in changing global climate on a short or long-term basis. When a large bolide smashes into the surface of the earth, it can send a massive amount of debris into the atmosphere. Some particulates are especially small and long lived, and will circulate the entire globe. These particulates affect climate in the same way that particulates from increased explosive volcanism do: they increase earth's albedo and lead to cooling. The degree of climate change depends on the type and amount of particulates and their duration in the atmosphere. Climate change can be regional or global, but is generally not long-lived.

How they all work together

It is clear at this point that any one of the above processes may impact regional climate and possibly even global climate in the short term. Major global climate changes require the coeval occurrences of multiple processes and the creation of positive feedback cycles. For example, as mountains grow the amount of exposed land surface increases, thus increasing albedo and weathering rates. Solar radiation is increased while simultaneously CO₂ is removed from the atmosphere. This may stimulate montane glacier formation, which further increases albedo, reduces sea-level and increases weathering rates. The processes feed off of each-other, ensuring that the trend towards global cooling continues until a powerful force interrupts it.

CHAPTER V: CLIMATE CHANGE ACROSS THE EOCENE-OLIGOCENE

Global

During the early Eocene, the global climate was very warm, forests blanketed Antarctica, and archaic taxa were scattered across marine shelves and continents (Cavelier et al., 1981). Diversification among pelagic organisms, such as protists and chordates, was at a maximum—from phytoplankton and dinoflagellates up to mackerel, sharks and whales (Cavelier, 1981). Global climate gradually cooled throughout the late Eocene, but experienced a marked acceleration towards colder temperatures at the Eocene-Oligocene. This is famously referred to as the “Greenhouse to Icehouse transition”. Major contributing factors include the opening of the Drake and Tasman Passages and subsequent isolation of Antarctica, as well as increased terrestrial weathering and reduction of atmospheric CO₂ and the various feedback mechanisms associated with each (Zachos et al., 1992; 1994; 1999; Lear et al., 2000; Exon et al., 2002).

The effects of global cooling in the marine realm included a sudden drop in seawater temperatures, and modified ocean chemistry due to ventilation and global changes in the oceanic carbon inventory—as evidenced from oxygen and carbon isotopes

and trace element analysis in biogenic carbonates (Cavelier et al., 1981; Miller, 1992; Diester-Haass and Zahn, 1996). The warm climate of the early Eocene, aided by orogenic events worldwide, promoted intense chemical weathering of siliceous rocks at high latitudes, resulting in high concentrations of dissolved silica in the oceans (Lear et al., 2000). Cooling stimulated upwelling of silica-nutrient rich water, and intense biosilicification occurred (Lear et al., 2000). The CCD also dropped up to 2,000 meters deeper, while simultaneously oceanic sedimentation rates declined and global sea levels fell (Cavelier et al., 1981; Kennett et al., 1985). Cooling at the Eocene-Oligocene also marked one of the largest extinctions of marine invertebrates in the Cenozoic period (Ivany et al., 2000).

The changes in temperature and ice volume are largely known from studying $\delta^{18}\text{O}$ records of the time period. Deep-sea benthic $\delta^{18}\text{O}$ records are very well constrained for the last 50 million years, having been obtained from ocean cores in various basins. The cores were collected as part of the Deep Sea Drilling Project (DSDP) or the Ocean Drilling Project (ODP), and have immensely aided our understanding of global climate change, see Figure 5. In the Pacific Ocean, DSDP Sites 167, 171, and 44 contain sedimentary records spanning the Eocene-Oligocene. Site 167 shows a $\delta^{18}\text{O}$ increase from zone P14-P17 by 1.08‰ in planktonic foraminifera, and 1.58 per mil from P12-P17 in benthic forams (Cavelier et al., 1981). At site 171, the $\delta^{18}\text{O}$ of planktonic forams increase 1.31‰ from zone P12-P13, while the benthic record increased by 1.17‰ (Cavelier et al., 1981). Planktonic foraminifera $\delta^{18}\text{O}$ increased by 1.5‰, and benthic by 1.25‰ at site 44 (Cavelier et al., 1981). Isotopic data from biogenic carbonates in the Pacific Ocean (near New Zealand and Australia) as well as the North Atlantic (near

northern Europe), show $\delta^{18}\text{O}$ increases by more than 1‰, corresponding to a temperature drop of at least 4°C (Cavelier et al., 1981).

Several DSDP projects were conducted in the Atlantic Ocean, which showed similar results. DSDP sites 198, 401, 116 and 366 clearly show an increase of $\delta^{18}\text{O}$ at the Eocene-Oligocene (Cavelier et al., 1981). However, there are frequent sedimentary hiatuses found at the boundary, which could be hiding the real extent of the isotopic shift, especially at sites 398, 401 and 116 (Cavelier et al., 1981). The shift in all Atlantic sites is approximately 1‰-1.5‰ (Cavelier et al., 1981). At DSDP Leg 90, Site 592 a distinct increase of ~1‰ at the Eocene-Oligocene boundary is noted (Kennett et al., 1985). Benthic and Planktonic foraminifera $\delta^{18}\text{O}$ values both show similar trends (Kennett et al., 1985). In the Aquitaine of the North Atlantic, the Biarritz (Cachaou) section, a coastal deposit, shows a +2.5‰ increase of $\delta^{18}\text{O}$ in the bulk carbonate analysis at the Eocene-Oligocene boundary (Cavelier et al., 1981).

The benthic $\delta^{18}\text{O}$ record of ODP site 689, on the Maud Rise in the Southern Antarctic Ocean, affirms other records of a shift towards more positive values in the early Oligocene (Miller, 1992; Diester-Haass and Zahn, 1996). $\delta^{18}\text{O}$ was analyzed by Shackleton and Kennett (1975) in the subantarctic, and the change was ~1.25‰ (Cavelier et al., 1981). Margolis et al. (1975) found a comparable shift in the isotopic composition of coccoliths (Cavelier et al., 1981). Additionally, the covariance between planktonic and benthic $\delta^{18}\text{O}$ suggest at least 0.3-1.0 per mil $\delta^{18}\text{O}$ increase is due to ice growth in the Earliest Oligocene (Miller, 1992).

The take-home message from these drilling projects is that the Eocene-Oligocene boundary exists at a threshold, where a large step toward more positive oxygen isotopic

composition of marine carbonate follows a gradual trend of enrichment throughout the late Eocene (Kennet et al., 1985; Miller, 1992; Diester-Haass and Zahn, 1996; Lear et al., 2000; Zachos et al., 1992; 2001; Ivany et al., 2003). This is commonly referred to as the “Oi-1 event” (Ivany et al., 2003).

The cause has been a point of contention. It was first attributed to a five degree Celsius temperature drop associated with the onset of thermohaline circulation, but more recently it has been linked to the rapid expansion of ice on Antarctica in the earliest Oligocene (Lear et al., 2000). Increased weathering from the growth of the Himalayas, coupled with the orogenesis of the Alps, and the increase in mid-latitude albedo due to sea level drop, further combined with a slowing of deep exchanges between the Atlantic Ocean and Tethys Sea, all aided in producing a major climate deterioration (Cavelier et al., 1981). Evidence supporting major glaciation at the Eocene-Oligocene boundary includes ice rafted debris, weathered clay mineral compositions, microfossil assemblages, and sea level changes and hiatuses (Cavelier et al., 1981; Miller, 1992; Zachos et al., 1992; 1999; Lear et al., 2000). The complex relationship that temperature and ice volume have on the $\delta^{18}\text{O}$ record complicates climate interpretations. It is important to sort out the influence of temperature and ice volume, as they both serve as indicators of climate and play an active role in shaping regional and global climatic conditions (Dwyer, 2000).

A high resolution Mg/Ca record at the Eocene-Oligocene boundary allows for the calculation of the seawater component ($\delta^{18}\text{O}_{\text{sw}}$) (Lear et al., 2000), allowing the temperature and ice volume signals to be unraveled. The Mg temperature record shows a consistent long-term decrease in deep-ocean temperature of $\sim 12^\circ\text{C}$ over the past 50 million years, and the curve is similar to the corresponding $\delta^{18}\text{O}$ record, with a major

cooling phase noted at the late Eocene-early Oligocene after a period of more stable temperatures (Lear et al., 2000). This independent deep sea temperature record allows for removal of the temperature effect from the benthic $\delta^{18}\text{O}$ record, thereby providing a record of the $\delta^{18}\text{O}_{\text{sw}}$ component which is largely a function of global ice volume (Dwyer, 2000; Lear et al., 2000). The data confirm that although temperature was decreasing, most of the signal results from the buildup of continental ice in the earliest Oligocene (34 My), and therefore each “step” in the benthic $\delta^{18}\text{O}$ record can be attributed to rapid accumulations of continental ice after threshold levels in the climate system were attained by gradual forcing mechanisms (Lear et al., 2000). It is clear that the periods of ice accumulation occur when deep water temperature decreases (Dwyer, 2000).

Paleotemperature trends are also supported by the location of the isotherms. Currently, the 18-20°C January isotherm is in the tropics, and can be found wherever reef-building corals, larger foraminifera, and mangroves exist in the marine environment (Cavelier et al. 1981). During the late Eocene this isotherm was located at latitude 50° North, and shifted to 45° north by the early Oligocene, as evidenced by the location of *Hantkenina*, *Globorotalia cerroazulensis*, and *Nummulites fabianii*. This would represent a temperature difference of 2.5°C in the present day surface waters in August (Cavelier et al., 1981).

Using the Pleistocene $\delta^{18}\text{O}$ sea-level calibration (0.11‰ per 10m) of Fairbanks and Matthews 1978, Miller (1992) determined that the $\delta^{18}\text{O}$ increase represents 30-90 m of glacioeustatic lowering. However, using the calibration of 0.055‰ per 10 meters of Miller et al. 1987 sea level may have been lowered by 180 meters (Miller, 1992).

Increased ventilation of the deep-sea is also indicated by seismic stratigraphy, hiatus distribution, and a large drop (up to 2,000 m) in the CCD (Miller, 1992).

The CCD fluctuates based on variations in carbonate distribution between deep and epicontinental seas (Miller, 1992) and temperature. The CCD decrease at the Eocene-Oligocene is compatible with the $\delta^{18}\text{O}$ evidence of sea-level drop noted by Miller (1992). The drop in the CCD was abrupt in the Pacific Ocean, but more gradual in the Atlantic and Indian Oceans (Kennett et al., 1985). DSDP data show that oceanic sedimentation rates, which had peaked in the middle Eocene, sharply decreased near the end of the Eocene and continued declining into the Oligocene (Cavelier et al., 1981). Sedimentary hiatuses were more common in the Atlantic Ocean than in the Pacific and in the southern Pacific than in the northern (Cavelier et al., 1981).

As global sea levels fell, tropical ocean gateways closed; concurrently, tectonic forces opened sub-polar gateways. Separation from Australia led to the Australia-Antarctic Gulf emptying into the Proto-Pacific Ocean, deflecting surface waters flowing south from the equator from reaching the South Pole (Exon et al., 2002). Antarctic originating surface currents no longer deflected north towards warmer climates, but travelled east, circling the new continent and creating the first circum-Antarctic current (Exon et al., 2002; Ivany et al., 2003; Barrett, 2003). The combined mechanisms thermally isolated the continent, creating a reduction in heat transfer with the equator and triggering deep water re-organization as cool surface waters sank resulting in deep water origination from the Antarctic, and the formation of the psychrosphere (Dwyer, 2000; Exon et al., 2002; Barrett, 2003; Ivany et al., 2003). Our current oceanic and atmospheric circulation patterns are rooted in the changes that occurred during this transition (Ivany et

al., 2003). There was no associated extensive glaciation of the once connected Tasmania or Australia, even though mountain glaciers already existed on Tasmania (Exon et al., 2002). Despite this and the fact that Tasmania was moving closer to the equator, evidence of cooling is found in the micro-fossils and clay minerals from that region (Exon et al., 2002). This trend continued into and throughout the Oligocene, creating a very dry Australia and extensive ice sheets over Antarctica, as the land bridge between the two regions completely subsided at the transition (Exon et al., 2002).

Longer lasting hiatuses in the western Atlantic are directly related to the deep-water reorganization event. Deep water originating in Antarctica contained high amounts of oxygen. The water was directed towards the western South Atlantic Ocean, where the increased oxygen levels led to greater rates of sediment erosion (Cavelier et al., 1981). Deep water reorganization is further supported by the deep-ocean $\delta^{13}\text{C}$ records.

During the late Eocene, $\delta^{13}\text{C}$ ratios dropped 0.5‰, and continued decreasing until the middle Oligocene to about 1‰ (Cavelier et al., 1981). There were variations in the amount of decline by region. Southern Ocean locations had higher $\delta^{13}\text{C}$ values than the Pacific Ocean during much of the late Paleocene and early Eocene, suggesting the Southern Ocean was proximal to a source of deep water (Miller 1992). $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ from 35.5-34.5 Ma indicates the coldest water was adjacent to Antarctica, and SCW was not apparent in the Atlantic and Pacific sites (Miller, 1992). The coldest waters may have mixed rapidly as they moved away from Antarctica (Miller, 1992). The $\delta^{13}\text{C}$ records and re-organization of deep-water circulation both support the dominance and importance of growing ice sheet on the Antarctic craton. Further support is gained from the Strontium (Sr) and Magnesium (Mg) records.

The Eocene-Oligocene boundary marks the most important break in the Mg and Sr distribution curves as Sr decreased and Mg increased diagenesis became the outstanding geochemical process for carbonates (Cavelier et al., 1981). While diagenesis of carbonates increased at the Eocene-Oligocene, carbonate sedimentation rates fell by the Oligocene (Cavelier et al., 1981). This is likely due to increased rates of terrestrial weathering and subsequent siliciclastic sedimentation, perhaps linked to the uplift of the Tibetan Plateau, one of the single largest increases in terrestrial weathering on earth.

Increased rates of weathering and subsequent cooling are further supported by an increased radiogenic Sr isotopic signal in marine records (Miller, 1992). The Sr isotopic record may actually be due to increased weathering on Antarctica instead of the Tibetan Plateau (Zachos et al., 1992; 1999). Weathering rates would have increased on the continent due to cooling that had already begun and the growth of the continental ice sheet (Zachos et al., 1992; 1999). Miller (1992) was able to reconcile the $\delta^{18}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ records and found that the Sr record increased soon after the $\delta^{18}\text{O}$ increase, which is consistent with glacial erosion of the Antarctic craton and increased continental Sr flux (Miller, 1992).

The changes noted in deep-sea temperatures, shallow seaway restrictions, changing ocean patterns, cooling atmospheric temperatures and increased seasonality led to worldwide pulsed turnovers in marine biota from the middle Eocene to the Oligocene (Ivany, et al., 2003). Algae decreased in diversity almost ubiquitously throughout the Eocene (Cavelier et al., 1981), as the meridional ocean surface temperature gradient between polar and subtropical regions steepened due to changing sea surface temperatures, the pattern of biological productivity paralleled this (Sloan and Barron,

1992; Zachos et al., 1994; Diester-Haass and Zahn, 1996). Many genera disappeared, including foraminifera, nannoplankton and dinoflagellates (Cavelier et al., 1981). Significant drops in Ostracode species richness at several sites in the Southern Ocean are noted (Cavelier et al., 1981).

Hundreds of warm and deep-water taxa became extinct and various planktonic faunal and floral taxa experienced decreased diversity (Kennet et al., 1985; Prothero, 1999). The formation of sea ice and subsequent heavy, cold water circulation along the ocean bottom may have been more important for marine organisms than global temperature (Cavelier et al., 1981). Deep water circulation may have caused the benthic flora and fauna upheavals noted in the geologic record (Cavelier et al., 1981).

Shallow marine invertebrates from the Northeast Pacific Bridge indicate stepwise temperature decreases from 3-4°C followed by unusual warming at 35 Ma, and terminal Eocene cooling of 6°C with pronounced seasonality of rainfall (Collinson et al., 1981; Ivany et al., 2003). Extinction was more important at the lower latitudes, while origination was more important at high latitudes (Ivany et al., 2003). Tropical regions were characterized by undiversified populations due to massive extinctions at the end late Eocene and beginning early Oligocene (Cavelier et al., 1981). Most groups show perturbations at or near the middle-late Eocene, and near or shortly after the Eocene-Oligocene boundary; but at the Eocene-Oligocene boundary taxa were affected indiscriminately (Ivany et al., 2003).

Nannoplankton and forams experienced similar evolutionary paths during this time (Cavelier et al., 1981; Prothero, 1999). Virtually no forms have a first appearance in the late Eocene-earliest early Oligocene, with the least diversification noted in the early

Oligocene; while at the same time there is no specific evolutionary trend in the planktonic forams (Cavelier et al., 1981). Genus extinctions of planktonic organisms began in the early late Eocene with heavier losses noted in the late Eocene (Cavelier et al., 1981). Only the genera *Turborotalia*, *Globigerina*, *Globorotaloides*, *Globigerinita*, *Cassigerinella*, *Chiloguembelina* and *Pseudohastigerma* cross the late Eocene-early Oligocene boundary, with *Chiloguembelina* and *Pseudohastigerma* disappearing in the early Oligocene (Cavelier et al., 1981). Most surviving early Oligocene planktonic foraminifera were small, low in diversity, and adapted to cold water (Prothero, 1999). Oligocene nannoflora was similarly impoverished compared with that of the middle to late Eocene (Cavelier et al., 1981).

In bathyal environments, the productivity regime and degree of oxygenation of the bottom are important factors in faunal turnover (Ivany et al., 2003). A shift from oxygenated, oligotrophic (low nutrient) water to oxygen limited, nutrient rich conditions, indicating cooling and upwelling is noted in bathyal and abyssal sediments at the Eocene-Oligocene based on benthic foram abundances and carbon isotopic work (Diester-Haass and Zahn, 1996; Ivany et al., 2003). Benthic foraminiferal assemblages experience pulses of extinction at the end of the middle Eocene, and/or at or close to the Eocene-Oligocene boundary in the Tethys and Paratethys in Italy, Israel, Circum-Alpine region, southern Australia, Caribbean and the US Gulf Coast (Cavelier et al., 1981; Ivany et al., 2003). Groups that were well represented in the late Eocene disappear just before the Eocene-Oligocene boundary (Cavelier et al., 1981). Only two species of the late Eocene group *Nummulites* continue to the early Oligocene; and only two first appearances are made—that of *N. intermedius* and *N. vascus* (Cavelier et al., 1981). According to Lear et al., the

argument for rapid ice accumulation with little to no temperature decrease is supported by the faunal distribution of benthic forams (2000). Benthic forams of this age show gradual, stepped extinctions from the middle Eocene to the late Oligocene, with no clear extinction event at Oi-1 (Lear et al., 2000).

Molluscs are the primary marine-shelf faunal component for which there are good data across the Eocene-Oligocene boundary because they are abundant, ecologically diverse, and well skeletonized (Ivany et al., 2003). The boundary between the Eocene and Oligocene is well marked in lacustrine and shallow-water molluscs in relation to temperature. Most forms became extinct in the late Eocene, some became extinct in the early Oligocene with a few species appearing in the early Oligocene, but molluscan turnover was greatest near the Eocene-Oligocene boundary (Cavelier, et al., 1981; Ivany et al., 2003). Similar molluscan extinction patterns are seen in the Gulf Coast, Atlantic Coast, Europe and the Pacific at the Eocene-Oligocene (Ivany et al., 2003).

At high latitudes in the North Pacific cosmopolitan molluscan taxa were strongly affected by extinction—more so than restricted taxa, likely due to the cosmopolitan ranges extending to warmer, lower latitudes (Ivany et al., 2003). Thermal tolerance was more important than the degree of endemism for survival at high latitudes; the Pacific Ocean groups were at higher latitudes than in other ocean basins, likely accounting for the greater turnover in the Pacific (Ivany et al., 2003). The preferential impact to warm water taxa strongly indicates that temperature is the dominant causal factor (Ivany et al., 2003).

Geographic distribution of European molluscs during the Eocene shows that after an almost exclusively south to north trending migration period in the early Eocene, the late Eocene was marked by a north to south migrational trend (Cavelier et al., 1981).

Mollusc shells analyzed by Buchardt (1978) from stratigraphic sequences in England, Holland, Germany, Denmark and Southern Sweden, show $\delta^{18}\text{O}$ from the middle Eocene to the early Oligocene increasing by 4‰. This is equivalent to a temperature drop of 10°C over a 2 million year period, though this may be “artificially” increased through variations in latitude position of the specimens, biologic fractionation, and the presence of aragonite (Cavelier et al., 1981). In northern Germany and Belgium, Oligocene molluscs are much less diversified than late Eocene molluscs—of 785 species in the late Eocene, only 75 transitioned to the early Oligocene (Cavelier et al., 1981). Along the Kamchatka Peninsula in Russia, a significant terminal Eocene molluscan turnover event is recorded, with warm water molluscan taxa suffering preferentially (Ivany et al., 2003).

Echinoderm diversity through the Eocene and into the Oligocene does not consistently follow that of molluscs or forams (Ivany et al., 2003). Nearly all regions decline in Echinoderm diversity from the mid Eocene to the Oligocene, but specific patterns differ between regions (Ivany et al., 2003). Pacific Northwest sand-dollars experienced a boundary event, as evidenced by their differences across the Eocene-Oligocene (Ivany et al., 2003). Carter (2003) suggests that the record of echinoid diversity across the shelf was primarily controlled by the relative proportion of carbonate versus siliciclastic sediment, and the balance between muddy and sandy substrates. Agreement has been found at sites in the North Pacific and Australia (Ivany et al., 2003). Oligocene echinoid fauna may have originated at higher latitudes, which is consistent with molluscs (Ivany et al., 2003). Some Echinoderm groups saw an increase in their diversity in the Oligocene (Cavelier et al., 1981).

Other marine invertebrate groups were impacted by climate change in the North Atlantic and Europe. At the Eocene-Oligocene boundary, significant drops in Ostracode species richness at several sites in Northwest Europe are noted (Cavelier et al., 1981). This is also associated with the transition from dominantly deposit-feeding to dominantly filter-feeding forms, indicating a trend toward low-oxygen conditions (Ivany et al., 2003). Several Tethyan marine vertebrates also went extinct at the Eocene-Oligocene, in some cases due to restructuring of food sources in response to cooling temperatures (Ivany et al., 2003). These include the archeocete whale *Basilosarus*, the giant sea snakes and the sawfish, *Propristis* (Ivany et al., 2003). All other marine vertebrate groups show either a gradual change or no change at all across the boundary (Ivany et al., 2003).

Changing continental climates are evidenced in the terrestrial biota. In the early Tertiary, there was little disparity of climate between vastly distant continents, and forests were present at high latitudes in both hemispheres (Sloan and Barron, 1992). This allowed for the migration of flora and fauna from Asia, south through Africa and across Antarctica into South America, and in the reverse direction (Briggs, 1987). The influx of East Asian vertebrate animals to Australia began in the mid-Tertiary (Briggs, 1987). The tropical zone extended down through South America, Africa and Australia, with tropical rainforest noted in Cameroon during the late Eocene (Jacobs, 2004). Early Tertiary Antarctica supported forest dominated by broad-leaved evergreens and evergreen conifers, species adapted to a high rainfall throughout the year and low to moderate ranges of temperatures; and warm-temperate fauna inhabited the shallow-waters of Australia and Antarctica (Briggs, 1987).

In the region where Antarctica and Australia separated, from the mid-Eocene through the Eocene-Oligocene transition, large shifts towards both warmer and cooler conditions were superimposed on a general cooling trend (Exon et al., 2002; Tripathi et al., 2005). Antarctica transitioned from a relatively warm environment to developing a continental ice sheet (Exon, 2002; Ivany et al., 2003; Barrett, 2003). Flora was not directly affected by cooling until the thermal threshold at the Eocene-Oligocene was reached; at which point a reduction in the extent of tropical rainforest is noted (Cavelier et al., 1981; Lear et al., 2000). There is evidence that winters became more severe, while summer temperatures were not affected as dramatically (O'Keefe, 1980; Sloan and Barron, 1992). A shadow effect could have been caused by increased atmospheric particulates/ash, due to the increase in explosive volcanism in the south Pacific/Indian Ocean region (Kennett et al., 1985). There were also major turnover events in terrestrial mammals, notably in Europe and North America. Difficulties in establishing good age control and correlating distant sections limits the ability to state that events were isochronous; however, the strong similarity of records between regions, environments and taxonomic groups suggests the turnovers were global (Ivany et al., 2003).

Additionally, major impact events, such as the Chesapeake Bay in North America and Popigai in Russia, may have had important effects on global climate change (Ivany et al., 2003; Cavelier et al., 1981). The enormous amounts of energy released by the two successive impact events may have induced the long-lasting perturbation of the global environment (Stoffler and Claeys, 1987). There are conflicting conclusions regarding the effect of Chesapeake Bay and Popigai Eocene impact events on climate (Ivany et al., 2003). Poag et al. (2003) believes that the impact debris played a role in sustained global

warming in the Eocene based on $\delta^{18}\text{O}$ from benthic forams immediately overlying the Chesapeake Bay impact. Additional evidence is provided by a negative global $\delta^{13}\text{C}$, possibly due to impact perturbations of the global carbon cycle (Poag et al., 2003).

Vonhof et al., 2000, use the 35.5 Ma Popigai impact event to contrast this argument using the $\delta^{18}\text{O}$ from the Popigai location to demonstrate accelerated global cooling. A notable peak in the dinoflagellate *Thalassiphora pelagica* at both locations indicates cooling and increased productivity (Vonhof et al., 2000). Since impact induced cooling is only apparent for a few years at most, some form of feedback mechanism, such as increased albedo, would be necessary and the impact could not be the sole source of cooling (Vonhof et al., 2000). Fawcett and Boslough (2002) note that cooling could have been re-enforced by the formation of a ring of impact debris around the earth, casting a shadow and shading the tropics for ~100,000 years (Fawcett and Boslough, 2002).

O'Keefe (1980) suggests that the tektite/microtektite strewn field across North America occurred at 34.4 my BP, and originated not as impact debris, but as extra-terrestrial forms which had formed a ring around the Earth. The ring's existence in the equatorial plane would create a shadow over North America in winter months, when the sunlight is directed toward the earth from the south, but not during the summer months, when the sunlight is directed from above (O'Keefe, 1980). This would account for winters becoming more severe, and the summers not being affected (O'Keefe, 1980). Sloan and Barron's (1992) climate model supports winter climatic sensitivity. However, it is more likely that these impact events only caused short term, possibly severe, regional effects on climate; they may have played a role in accelerating whatever underlying forces were already at work (Ivany et al., 2003; Cavalier et al., 1981).

It has been suggested that the late Eocene bolide impacts may have played a role in extinction at the time (Poag et al., 2003). Due to poor preservation, this issue is unresolved in the Carolinas, and shown to have no correlation in the Gulf Coast (Ivany et al., 2003). Given that so many researchers have noted extinction events beginning in the middle Eocene and ending in the early Oligocene (both well before and after the impact events), it is unlikely that Chesapeake Bay or Popigai had a significant effect on global biota (Ivany et al., 2003). The microtektites in the mid-late Eocene are the only association of an extraterrestrial and extinction event, and it is the least significant extinction event in the late Eocene (Swisher and Prothero, 1990).

Eurasia

Much of Europe and Asia reside at the same latitudes as North America, many of the depositional environments were therefore affected by similar climatic conditions. Comparisons between the regions are therefore useful for understanding the impact of global changes in climate. The orogenesis of the Alps and the simultaneous reduction of the Tethyan Sea exaggerated the climatic zonation of the continental and marine regions of Europe (Cavelier et al., 1981). The exaggerated relief of the Alps may have increased the albedo effect while the reduction of the Tethyan Province diminished the east-west circulation of warm surface water (Cavelier et al., 1981).

Several varieties of relict flora existed across Greenland, Iceland, Europe and North America in the early Eocene, including Ginkgo, Sequoia, Liriodendron and *Ceridiphyllum* (Briggs, 1987). Terrestrial tropical flora retreated in Europe from the Paleocene into the Eocene (Cavelier et al., 1981). Major extinctions of forest adapted

archaic species also occurred (Hartenberger, 1988) and were replaced with more modern fauna with a size distribution suggestive of dry, open rangeland (Retallack, 1992), but clear evidence of a drying trend is not noted (Prothero, 1994). Changes in European floras were subtle, likely because climate was moderated in many parts of Europe by location on humid coastal islands (Prothero, 1994). In general, the pollen record demonstrates increases in temperate plants and conifers and a loss of the last tropical and subtropical vegetation as late Eocene evergreens, bald cypresses, and reed marshes were replaced by mixed deciduous/evergreen flora (Prothero, 1994).

On the Isle of Wight, southern England, palynological evidence from the Hampshire Basin indicate an onset of cooling in the latest early Eocene with no evidence for a sudden climatic change at the end of the Eocene (Collinson et al., 1981). Palynoflora of the middle Eocene underwent gradual replacement of aquatic macrofloral assemblages by ferns while previously abundant tropical and subtropical macroflora were reduced to absence by the early late Eocene (Collinson et al., 1981). Contrary to this finding, the inland Paris Basin floral community pattern of evolution indicates a marked change in the late Eocene following the gradual diminution of tropical species to more temperate species through the Paleocene and into the Eocene (Cavelier et al., 1981). The average number of species increased from 90-270 in the late Eocene, quickly dropping to 50 across the transition (Cavelier et al., 1981).

An important break in the faunal succession of land mammals of the European Paleogene is known as the Grande Coupure (Cavelier et al., 1981). It was established primarily on the evolution of larger mammals between the late Eocene and late early Oligocene in Western Europe, although it is also noted in North America and Asia

(known as the Mongolian remodeling) (Cavelier et al., 1981; Hartenberger, 1998; Meng and McKenna, 1998).

North America

The US Gulf Coastal Plain is one of the best sections of fossiliferous Paleogene sediments in the world (Ivany et al., 2003). Precise location of the Eocene-Oligocene transition in the Gulf Coastal Plain vertical sections has been a matter of debate. The contact of the Red Bluff Clay with the underlying Shubuta Clay Member of the Yazoo Clay is typically defined as the lithologic boundary between the Eocene and Oligocene (Fluegeman, 2009). Many researchers recognized the Eocene-Oligocene boundary at the Jackson-Vicksburg lithologic contact in the western Gulf Coastal Plain, while others dispute this (Fluegeman, 2003). Detailed studies of the planktic and benthic foraminifera from the Mossy Grove core show that the Eocene-Oligocene boundary is actually within the Yazoo Clay, rather than at the Yazoo-Red Bluff lithologic contact (Fluegeman, 1996, 2003, 2009). In eastern Mississippi, the Eocene-Oligocene boundary is recognized through the use of planktonic foraminifera and calcareous nannofossils as occurring at the unconformable sequence boundary of the Yazoo-Red Bluff contact (Fluegeman et al., 2009). In Alabama, the Eocene-Oligocene boundary is placed within a condensed section at the contact between the Shubuta Member of the Yazoo Clay and the Red Bluff Clay (Fluegeman et al., 2009). The unconformity at the Yazoo-Red Bluff contact was produced by a fall in relative sea level possibly associated with Oi-1 (Fluegeman, 2009; Fluegeman et al., 2009).

The Jackson benthic foraminiferal fauna is typical of the Jackson Group, and spans the upper Eocene and lowermost Oligocene, while the Vicksburg fauna occurs exclusively in the lower Oligocene Vicksburg Group; thus the Eocene-Oligocene biostratigraphic boundary based on planktonic foraminifera in the Gulf Coastal Plain occurs below the benthic faunal transition (Fluegeman, 2003). Specifically, the Eocene-Oligocene boundary is 16.6 meters below the Jackson-Vicksburg lithologic contact—well within the Yazoo Clay, with no Vicksburg species identified in any of the samples at the Mossy Grove core (Fluegeman, 1996, 2003). The transition between the two benthic faunas is gradual, with some noticeable changes, namely last appearances of some forams prior to the Red Bluff Clay coeval with some first appearances (Fluegeman, 2003). Species distributions, however, show a sharp transition, with Jackson species having their highest occurrence at the Yazoo-Red Bluff contact and Vicksburg species having their lowest occurrence at or near the Yazoo-Red Bluff contact (Fluegeman, 2009).

Concurring evidence is found in the Mobil #1 Young, WCLC, and Ketler cores in southeast Mississippi. The benthic foram turnover occurs above the unconformity associated with the base of the Vicksburg Group (Fluegeman et al., 2009). Paleomagnetic studies conducted by Echols et al. (2003) on the Young core indicate that all but the upper 0.15m of the Yazoo Clay are within Chronozone 13r while all but the lower 0.46m of the Red Bluff are within Chron 13N (Fluegeman, 2009). Therefore, where samples were viable, the Red Bluff is Oligocene in age (Fluegeman, 2009). This indicates that the Jackson-Vicksburg benthic foraminiferal faunal turnover is not related to an event at the Eocene-Oligocene, but to an early Oligocene event, possibly the Oi-1 Antarctic glacial event (Fluegeman, 2003).

The Eocene-Oligocene marks the end of the characteristic shallow-marine Tethyan fauna in North America (Ivany et al., 2003). The Atlantic Coastal Plain molluscan diversity is characterized by pulses of extinction rather than origination (Ivany et al., 2003). US Atlantic Coast Molluscan faunas shared a taxonomic affinity with Gulf Coast assemblages (Campbell and Campbell, 2003). Both experienced similar step-wise extinction patterns across the middle to late Eocene, with no taxa from one stratigraphic unit surviving to the next (Campbell and Campbell, 2003). The Gulf Coastal Plain molluscan species experienced significant losses. More than 95% of the Gulf Coast molluscan species went extinct (Ivany et al., 2003), including 97% of gastropod species and 89% of bivalve species (Prothero, 1999). The Gulf Coast molluscan extinction correlates with type one sequence boundaries indicating sea level drop (Ivany et al., 2003; Fluegeman, 2009). In many cases the extinctions at the Gulf Coast were true extinctions rather than emigrations (Dockery and Lozouet, 2003).

New species originations with distinctly new biota were concomitant with extinctions throughout North American oceans (Ivany et al., 2003). First appearances must have included a substantial amount of immigration rather than evolution in situ, due to rapidity of originations and lack of severe diversity loss (Ivany et al., 2003). Echinoderm diversity in the Southeast actually increased from the middle to late Eocene, and then declined into the early Oligocene (Carter, 2003). In Alaska, origination exceeded extinction, and Alaskan fauna were unaffected by Eocene-Oligocene boundary extinction (Ivany et al., 2003).

The backfilling of the rise of the Cascadian Province and Rocky Mountains is expected to have had a similar effect as the Alpine orogenesis and closing of the Tethys

Sea in Europe, and the Tibetan plateau uplift in Asia – exaggeration of the climatic zonation between the continental and marine regions of North America. Increased land surface led to higher albedo and increased weathering rates, which contributed to removing more CO₂ from the atmosphere. Reduction of the intercontinental seaway diminished longitudinal circulation of warm surface water. Volcanic activity also increased in the region, stacking sedimentary sequences with ash. Ejected particulate matter would have added to the albedo effect, though this may have been countered by the concomitant release of greenhouse gases.

Rising mountain ranges would have interfered with regional air current patterns, creating rain shadows and microclimates for parts of North America. Based on topography-atmospheric circulation models conducted by Sloan and Barron, variations in the height of mountain elevations on Eocene continents had the greatest effect on precipitation distributions at the windward (upwind) side of the mountain ranges—especially in North America (Sloan and Barron, 1992). The result was a decrease in precipitation for the continental interior due to the rain shadow effect as mountain height increased (Sloan and Barron, 1992). Orographically induced drying of continents could have begun when mountains were as low as 1000 meters (Sloan and Barron, 1992).

Cooling and concomitant drying are seen in paleosols across the continent. Paleosols in the Pacific Northwest are represented by 500 ky to 1.6 My of gradual climate change and concomitant soil development recording a transition from a subtropical to a humid temperate environment at 34 Ma corresponding to the marine event Oi-1 (Bestland et al., 1997). Stepwise increases in central Oregon sedimentation rates across the Eocene-Oligocene reflect a drying climate and sparser landscape cover (Bestland et al., 1997).

In Wyoming, floodplain deposits indicate climate transition from wetter to drier conditions, and were ultimately replaced with eolian deposits, indicating increased aridity (Kraus, 1997; Prothero, 1999). Even in the Big Bend region of west Texas where the climate of the earliest Oligocene was likely tropical with high rain fall, a pronounced (2-3 month) dry season is noted (McBride et al., 1968).

The debate regarding the role of climate change on North American flora and fauna at the Eocene-Oligocene is ongoing (Hartengerber, 1998). In addition to paleosols and sediments, climate change throughout the continent is largely recorded through the extinction of Eocene forms adapted to local conditions and their replacement by immigrants in the early Oligocene (Cavelier et al., 1981). Generally, the floral fossil record supports drying associated with cooling throughout the Eocene and across the Eocene-Oligocene boundary. Much of North America experienced a floral change from paratropical forests, which were widespread from the late Cretaceous into the early Eocene, to broad-leaved deciduous forests by the Eocene and into the Oligocene similar to those in New England today (Cavelier et al., 1981; Kraus, 1997; Prothero, 1999). Eocene vegetation in the Willwood formation indicates a mean annual temperature of 13°C-18°C and seasonal rainfall (Kraus, 1997). Pollen from San Diego and the Gulf Coast marine sequences show that cooling and drying actually began in the middle of the middle Eocene; tropical palms, elms, and walnuts were replaced by sycamore and hickories, and shrubs, herbs, and grasses (Prothero, 1994). The end of the middle Eocene shows further cooling and drying as oak pollen counts increase (Prothero, 1994). The pattern of evolution of the Northern Hemisphere flora leads to the conclusion that at the

Eocene-Oligocene boundary there was a general drop of temperature accompanied by a period of great drought (Cavelier et al., 1981).

Gulf Coast Jackson formation floras represent tropical rain forest with MAT of 28°C (82°F), but the overlying Vicksburg formation contains too few megascopic plants for comparison (Prothero, 1994). Changes in pollen are striking, though. In Mississippi and Alabama, massive extinctions of late Eocene pollen taxa are noted, while the Vicksburg formation includes abundant cool, dry-adapted oak pollen (Prothero, 1994).

Declining diversity among turtles, crocodiles and champsosaurs in the latest Tertiary are indicative of mild winters with approximately 13°C cold month mean temperatures, and dry climates (Hutchison, 1982). Aridity may have been annual or seasonal, resulting in few permanent bodies of water (Hutchison, 1982). Mammalian evolution through the Eocene was characterized by immigration as well as extinction. Asian animals had been migrating to North America periodically via the Bering land bridge since the Cretaceous, but in the late Eocene mammals adapted to woodland savanna and scrubby habitats arrived (Briggs, 1987). Immigrants included several groups of lophodont rodents, pig-like enteledonts, several families of artiodactyls, and several groups of perissodactyls, including tapiroids and rhinocertoids (Briggs, 1987). There were numerous insect migrations back and forth across Beringia (Briggs, 1987). One interesting example is the primitive generic line *Sortosa*, of the caddisflies, which originated in Asia, migrated to North America in the late Cretaceous, then migrated to South America and produced a descendant line, *Chimarra* (Briggs, 1977). *Chimarra* migrated to North America, then back to Asia in the Paleocene-Eocene, then finally back to North America in the late Eocene (Briggs, 1987).

The Terminal Eocene is represented by relatively minor extinctions in the Chadronian-Orellan North American Land Mammal Age (NALMA) compared to the middle Eocene (Swisher and Prothero, 1990). Most of the faunal responses during the Chadronian-Orellan transition had already taken place more than 250 ky before the climatic crash of the early Orellan (Prothero, 1999). That being said, the Grande Coupure event in mammals, rodents and reptiles of Europe is noted in North America (Cavelier, Claude et al., 1981). This was a period of low diversity, followed by extinction and subsequent renewal of equally low diversity (Cavelier, Claude et al., 1981). The brontotheres, oromerycid artiodactyls, and the cylindrodont rodents disappeared, and the oreodont *Miniochoerus* underwent slight dwarfing (Prothero, 1999).

The White River Badlands

Badlands National Park, South Dakota USA is located in the southwest section of the state. The modern average low temperature is 37 F (2.7°C) while the modern average high temperature is 63 F (17.2°C), with MAP of 15.6 inches. The climate is generally characterized by hot, dry summers (July and August temperatures ~91.5 F, (33°C) with precipitation of ~1.7 inches) and cold, dry winters (December and January temperatures are 36.5 F (2.5°C) precipitation is .295 inches) (nps.gov). Sudden storms can occur in any season, although most precipitation occurs in the summer months (nps.gov). Vegetation is currently defined as a mixed-grass prairie, bridging the gap between the moister tall-grass prairie to the east and the drier short-grass prairie to the west (nps.gov). Some cattle and bison herding is conducted on and around the park. Prairie dog populations are abundant, and mountain goats are also common in the park. Much of the foundation for

the current vegetative and faunal adaptations to the land are due to the transition in the early Oligocene, as will be discussed.

At the Pinnacles area, at least ten fossil soil series can be recognized, each representing a particular ancient environment (Retallack, 1983). Additionally, the Badlands region has always been among the most important collecting areas for White River mammals, with serious study beginning in the 1940's (Prothero and Whittlesey, 1998). These were extensively discussed in the preceding chapters.

“Smectite and the presence of fossil alligator and soft-shelled turtle at comparable stratigraphic horizons elsewhere in South Dakota are evidence that for a long time, climate in this area was humid and subtropical to warm temperate” (Retallack, 1983). Abundant rainfall and a forested subtropical environment are evidenced throughout the region and land snails adapted to wet, subtropical habitats were abundant (Prothero, 1999; Stoffer, 2003). It is important to note that Hackberry endocarps are the only direct fossil evidence of a plant formation, even fossil pollen samples are not found, as they are poorly preserved in the Badlands (Retallack, 1983; Prothero, 1994).

The climate became drier as the land steadily continued to rise and fill with volcanic ash from the mid to late Tertiary; by the end Tertiary the region was a desert with eolian deposition (Stoffer, 2003). A subtle change in vegetation from forests in the yellow mounds to woodlands during deposition of the Interior paleosols, and finally open woodland during the Gleska series is observed in the paleosols (Retallack, 1983). The land snails were replaced by smaller taxa with restricted apertures, typical of drier climates; while aquatic reptiles and amphibians gave way to dry land tortoises (Prothero, 1999). Savanna-adapted fauna likely migrated into the Great Plains from elsewhere

before and during the deposition of the Chadron because they appear during a time of extensive woodlands in the Chadron (Retallack, 1983). The woodlands of the Eocene gave way to savannah in the Oligocene (Stoffer, 2003).

While the Eocene-Oligocene boundary is defined stratigraphically as the Chadron-Brule contact, this contact cannot be used to define the Chadronian-Orellan NALMA in the Badlands region because the contact is time transgressive (Prothero and Whittlesey, 1998). High resolution Ar^{40}/Ar^{39} dating and comparison to the magnetostratigraphy record for the Badlands National Park imply that the Chadronian-Orellan NALMA faunal transition corresponds to the terminal Eocene event and at the division between Chron C13R and C13N, 33.7 Ma (Swisher and Prothero, 1990; Prothero and Swisher, 1992; Prothero and Whittlesey, 1998) (Figure 4). However, the Chadronian-Orellan NALMA must be definable on the basis of fossil taxa.

The Chadronian-Orellan boundary is subtle, with no major “faunal break” (Prothero and Whittlesey, 1998). In the Badlands region the most notable change in fauna is that mammal remains are much more abundant in the Brule, and the once prominent Chadronian titanotheres are missing (Prothero and Whittlesey, 1998). The most useful taxa for characterizing the Chadronian-Orellan transition are the First Appearance Datum (FAD)'s of *Hypertragulus calcaratus*, *Leptomeryx eransi*, *Palaeolagus intermedius* and *Miniochoerus chadronensis* and the Last Appearance Datum (LAD) of *Poebrotherium eximium* and *Miohippus grandis* (Prothero and Whittlesey, 1998). Prothero and Whittlesey recommend the beginning of the Orellan NALMA be defined by the FAD of *H. calcaratus* due to its sudden appearance and ease of recognition and characterized by

the rest of the taxa commonly found in the section and their first and last appearances (Prothero and Whittlesey, 1998). These taxa are described below.

Though faunal change was subtle, extinctions and originations did occur across the Eocene-Oligocene. Extinctions and reductions in diversity at the Chadron-Brule contact were likely due to a disruption in the supply of water and mean annual temperature drop (Hutchison, 1982). Local and temporary steppe environments “probably provided important selection pressures for the evolution of prairie-adapted faunas of the late Tertiary” (Retallack, 1983). In the late Chadronian, the common Badlands rhinoceroid, *Subhyracodon occidentalis*, first occurs and shows no noticeable change in the Orellan (Prothero and Whittlesey, 1998). The amynodontid rhinoceroid, *Metamynodon planiformis*, is the basis of the name of the Metamynodon channels in the lower Scenic Member (Prothero and Whittlesey, 1998).

Several brontothere species also existed from the Chadronian through the late Orellan (Prothero and Whittlesey, 1998). The *Miohippus grandis* ranges only from the early to late Chadronian in the Big Badlands, making its LAD an indicator of the NALMA (Prothero and Whittlesey, 1998). The canid amphicyonids, or “bear dogs” *Daphoenictis* last appears at the end Chadronian, and *Paradaphoenus* first appears in the Orellan (Prothero and Whittlesey, 1998). *Daphoenus vetus* and *D. hartshornianus* range from the late Chadronian through the early Orellan (Prothero and Whittlesey, 1998). Cat-like mammals, *Dinictis felina*, range from the Mid-Chadronian to the late Whitneyan, while *Pognodon platycopis* occurs in the Orellan of South Dakota and the Whitneyan of South Dakota and Nebraska (Prothero and Whittlesey, 1998). *H. occidentalis* occurs in the late-early Orellan of South Dakota (lower nodules), North Dakota, Nebraska, and

Wyoming and the Whitneyan of South Dakota (Prothero and Whittlesey, 1998). *H. sicarius* is found in the lower nodules of the badlands (Prothero and Whittlesey, 1998).

In the Orellan, the most common mammals, next to the Oreodonts, were the small, deer-like artiodactyls *Leptomeryx* and *Hypertragulus* (Prothero and Whittlesey, 1998). The canid *Mesocyon temnodot* first occurs in the Orellan of South Dakota and the Whitneyan of Nebraska and Colorado (Prothero and Whittlesey, 1998). A smaller carnivore, the *Drassonax harpagops*, is found in the Scenic member of the South Dakota Badlands (Prothero and Whittlesey, 1998). Death assemblages accumulating under woodlands during early-late Oligocene were mainly Oreodon (*Merycoidodon*, 30%) and horse (*Meshippus*, 24%) (Retallack, 1983). Under savannas of interfluves, it was mainly rabbits (*Paleolagus*, 26%), the deer-like *Leptomeryx*, 21% and *Hypertragulus*, 17% (Retallack, 1983). Both were littered with tortoise shells of *Styemys* (Retallack, 1983).

The middle-late Oligocene is marked by the deposition of the Poleslide member of the Brule Formation. The ecology consists of meadows, savanna woodlands and swampy savanna (Retallack, 1983). The fossil record shows a change in assemblages of land snail, amphibian, and reptile fauna, including a marked decrease in alligator body size and numbers—indications of a cooler and drier climate (Retallack, 1983; Evans and Welzenbach, 2000). The general trend in the Poleslide is increased aridity and less vegetation than Eocene deposits; characterized by streamside early successional vegetation and gallery woodland, with more open wooded grassland away from streams (Retallack, 1983; Retallack, 1992). Savanna-adapted mammals were well suited to the savanna and woodland vegetation of the Brule (Retallack, 1983).

Higher in the Brule and into the Sharps Formation, an increase in abundance and diversity of burrowing mammals, decreasing body size, and dentition approaching subhypsodont is seen; this indicates a drier, cooler, less productive ecosystem (Retallack, 1983;1992). Although extensive steppe or prairie in the Sharps formation is supported by mammal fossils high diversity, and persistence of the “White River Chronofauna” elements (including long necked camels, and lack of hypsodont dentitions) are indications that the fauna were not as adapted to the steppe as Pliocene and modern faunas (Retallack, 1983). The landscape of the Sharps Formation was dry with elevated and extensive grassy floodplains dissected by wooded stream gullies and their associated terraces (Retallack, 1983).

In summary, the climate in the Badlands, SD region appears to have transitioned from humid and warm, to temperate and sub-humid in the early Oligocene, to arid by the middle Oligocene. This has been deduced from studying the sediment and paleosol compositions, and faunal and floral fossil evidence. Retallack’s evidence and identification of paleosols confirm climatic changes; however the interpretations lack precision because climate is only one of a number of soil-forming factors (Retallack, 1983). Vegetation transitions are best summarized as changing from moist forests of 38 Ma to dry forests of 34 Ma to dry woodland by 33 Ma to wooded grassland, with streamside gallery woodland by 32 Ma, and large areas of open grassland by 30 Ma (Retallack, 1992). Faunal assemblages showed decreases in reptilian and amphibian fossils, and other animals adapted to wet, subtropical habitats, with land mammal extinctions occurring to a lesser extent across the Eocene-Oligocene boundary (Prothero,

1999). Early Orellan assemblages included animals better adapted to drier climates, including land tortoises (Prothero, 1999).

Zanazzi et al (2007) looked at in vivo stable isotope compositions of fossil tooth enamel and diagenetic stable isotope compositions of fossil bones to derive a high-resolution (~40,000 year) continual temperature record over the transition in the Badlands region of the United States. Tooth enamel isotopes are expected to reflect the dietary conditions in the $\delta^{13}\text{C}$ record, and isotopic composition of meteoric waters through the $\delta^{18}\text{O}$ record, with seasonal changes recorded in $\delta^{18}\text{O}$ zoning along the length of each tooth (Zanazzi et al., 2007). They found a large drop in Mean Annual Temperature (MAT) of $8.2 \pm 3.1^\circ\text{C}$ over 400,000 years, small increases seasonality and virtually no change in aridity (Zanazzi et al., 2007).

Given that the depositional environment of the Badlands is entirely derived from qualitative evidence, except for one study, there is pressing importance to assemble quantitative records from which to draw conclusions. North American continental climate data are ambiguous at best- indicating changes in MAT from $0\text{-}8^\circ\text{C}$ (Zanazzi et al., 2007). The timing between continental and oceanic climate change is also unclear (Zanazzi et al., 2007).

The isotopic record obtained from Badlands paleosols can serve as a direct comparison to the tooth isotope study presented by Zanazzi et al. (2007). The paleosol $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records can also be compared to the marine deep-sea record as well as Gulf Coast shallow-marine isotopic records through magnetostratigraphic correlations.

CHAPTER VI: CURRENT FINDINGS

This research establishes a continental interior isotopic record from paleosol carbonates. By comparing the interior continental record with various marine records, it is possible to begin deciphering the role of global climate change on the mid-continent, the differences or similarities in timing between the two realms, and whether or not decoupling between mechanisms was taking place. The role of regional influences can be assessed. The Badlands isotopic record can also be compared to the qualitative climate records either validating or undermining them.

Field Methods

A site visit to Badlands National Park, South Dakota, was conducted from 27 October, 2005 through 31 October 2005. Field samples were collected on the 28th-30th of October, 2005. Dr. Rachel Benton, park Geologist, assisted in pointing out the location of three sections of most interest—the Conata Picnic Area, Chamberlain Pass, and the Retallack Type Section (Figure 6). Each of these locations contained an excellent exposure of Eocene-Oligocene paleosol stratigraphy and was chosen due to their proximity to other well-documented sections in which the stratigraphy and paleomagnetic record are well understood. Retallack (1983, 1985, 1992), Swisher and Prothero (1990),

Prothero and Swisher, (1992) and Prothero and Whittlesey, (1998) provide detailed lithologic and magnetostratigraphic descriptions for these sections. This allows for the samples collected to be tied to a stratigraphic timeline anchored in paleomagnetism. The isotopic record can thus be compared to other records of the same magneto-chronology.

The Badlands National Park requires special permits for collecting rock or fossil samples. The park is well known for spectacular mammal fossil finds, particularly at the Pig Dig near the Conata Picnic Area. Illegal fossil collecting has been a problem, requiring the park to maintain strict control over collection procedures. Under the rules of the permit only hand specimen or smaller sizes of rocks and soil were allowed for collection. Care was taken to sample off the road, in an inconspicuous area, in order not to encourage fossil poaching. Additionally, the southern portion of the park is co-owned by the Oglala tribe in the Lakota Sioux, of the Pine Ridge Reservation. Additional permits are required to sample that portion of the park. Such permits were not sought after, as there were multiple good exposures in the northern section of the park.

Paleosols in this study area were recognized as brown, pink, grey, grey-green, and white horizons, and generally were well correlated to the stratigraphy of Retallack (1983). The same sampling procedure was used throughout the investigation. GPS coordinates were recorded for each sampling location using a Garmin II personal GPS device, with an accuracy of plus or minus (+/-) 5 meters. Each location was observed, measured and sampled above and below the Eocene-Oligocene contact. Stratigraphic thicknesses were measured by use of a 1.5-meter Jacob's staff. Each layer that was accessible was sampled.

Samples were collected by digging into the exposure to collect paleosols that were not exposed on the surface to ambient environmental conditions. The stratigraphy was generally not well lithified. Paleosols were typically found to be hand specimen size, loosely in place. This resulted in generally porous conditions for the exposure as a whole, but lab results do not indicate modern climatic overprints, as will be discussed later. Lighting plays an important role at the park, as the angle of the sun affects the color of the rocks and soil. Coloring will change throughout the day, sometimes significantly—a grey rock can appear red and vice versa. Care was taken to note the sample color in the field, and to re-evaluate sample color under incandescent lighting as well.

Field notes were taken to document location, color, thickness, mineral composition, and grain size where applicable. For samples noted as paleosols, the grain size was too small to be seen with a hand lens. Mineral composition was determined by field examination of hand specimens and the use of hydrochloric acid. Where a reaction occurred on the sample after applying hydrochloric acid, it was determined that the sample contained calcite, and was flagged for lab analysis. Samples were placed in individual Ziploc bags labeled with their sample number and location. Stratigraphic columns were drawn in the field for each sampling location (Figures 7-10).

The first location sampled was the Conata Picnic Area, (Figures 6-8, 11, 12). This sampling location was located past the Pig Dig in a stream cut bank near a site where significant paleomagnetism research was previously conducted by Prothero and Whittlesey (1998). The specific location is located at UTM Coordinates 13T 0725663 Easting and 4856976 Northing (Lat 43.831, Long -102.193), with an elevation of 2,599 feet above sea level. Two measured sections were recorded from this location, 5 and 3

meters in length (Figures 7 and 8). A total of 22 lithological units above and below the contact were found. Within these units were 13 distinct calcite rich layers. A total of 22 samples were collected, 14 from calcite rich layers, with 12 being sent to the lab for analysis. Of those sent to the lab, five were from the first measured section, while 7 were from measured section #2.

Chamberlain Pass, located off highway 44 in the “dog-leg” section of the park, was the second sampling location (Figure 3, 9, 13). It is approximately ½ mile after the park entrance on the north side of highway 44, and is often referred to as the type Scenic section. There are some good bluff exposures in this area that could be climbed to reach the Scenic member of the Brule Formation. The GPS location is located at 13T 0701404 Easting 4847065 Northing +/- 5 meters (Lat 43.749, Long -102.498). One measured section, 12.5 meters long was recorded at this location (Figure 9). A total of 27 lithological units were observed, 30 samples collected, 14 of which were calcareous and sent to the lab.

The final locality was at the Retallack Type Section, located in Dillon Pass (Figure 3, 10, 14). This is where Retallack collected samples he identified as paleosols from the Eocene-Oligocene transition. A GPS location was taken at the mouth of the canyon, the location is at 13T 0723296 Easting, 4859131 Northing, +/- 6 meters (Lat 43.851, Long -102.222). Sampling actually occurred about halfway into the canyon, and then along the vertical exposure. One measured section 10.5 meters long was recorded (Figure 10). Fifteen lithological units were noted, 16 samples collected, of which 6 were calcareous and sent to the lab for further analysis. Lighting in this area was particularly variable throughout the day, making identification of stratigraphic layers very difficult.

Lab Methods

Samples that contained calcite were taken to the Indiana University Department of Geology Stable Isotope lab for oxygen and carbon isotopic analysis, September 4th 2007. Doctoral candidate Melanie Everett and lab director Dr. Peter Sauer assisted with sample preparation and conducted the isotopic sampling procedures. A total of 31 samples were selected for analysis (Table 1).

A Thermo Finnigan GasBench II Mass Spectrometer was utilized for isotope ratio determination of the samples. The GasBench II analyzes small gas samples, either CO₂ or N₂, between 200nmol and 20mmol total sample size. In this study the gas, CO₂, was released from solid phase paleosol carbonates into the headspace of the sample vial via the sample preparation methods described below, then passed through a Valco sampling port using a stream of Helium (He) and Arsenic (As). In this way, ¹⁸O and ¹³C were simultaneously measured from the samples.

Five samples were initially selected to determine the optimal weight and sampling time necessary for further analysis based on calcite concentration. The five initial samples had a range of calcium carbonate concentrations. The preparation method is as follows:

- 1) Samples were ground using mortar and pestle.
- 2) 1.50 ml volume of sample was placed in a small flat-top capped graduated centrifuge vial to allow for multiple runs.
- 3) The mortar and pestle were decontaminated between each sample with 1 rinse of hydrochloric acid (HCL), 3 rinses of de-ionized (DI) water, and 1 rinse of ethanol.

- 4) Each sample was divided among 4 round-bottomed glass vials according to weight.
 - a. Used approximate weights of 0.5, 1.0, 1.5 and 2.0 mg.
- 5) Twenty round-bottomed glass vials were prepared with reference material at weights of approximately 0.5 mg.
 - a. Ten samples used the reference material NBS19 (National Board of Standards).
 - b. Ten samples used the reference material Corydon Limestone (internal standard).
- 6) All samples were placed in a tray in a vacuum oven dryer at 85-100°C and 25-30mm Mercury (Hg) for approximately one hour.

A vacuum line to remove headspace from the sample vial was attempted after using the dryer, though it was unsuccessful. The method for applying vacuum follows. First, a plastic cap with a septum is fitted over the glass vial, then a glass blown valve with a fitted hypodermic needle is placed on top of the septum. The valve is connected to the vacuum line and pressure is removed by piercing the septum with the needle. The vacuum is applied until the thermocouple vacuum gauge reads <50 millitor (micrograms of Hg). This method was unsuccessful in achieving the necessary pressure. The traditional flushing method using the GasBench II and a 2-holed needle was then implemented. This method involves loading the sample tray into the GasBench II and inserting a 2-hole needle into the headspace of the sample vial during sampling. The top hole in the needle removes air in the headspace; while the bottom hole delivers He and As. Flushing took place for six minutes.

Using the autosampler and a double needle set up, the sample gas is analyzed. The double needle set up applies phosphoric acid (H_3PO_4) to one row of sample vials, while the other needle samples the adjacent row. The application of H_3PO_4 liberates the sample gas from the carbonate. The block holding the sample vials was heated to 72°C to speed the reaction between carbonates and H_3PO_4 . For the first run, the acid needle calibration was set to 8 strokes per drop. The reaction interval was set at two hours. This ensured equilibration between the paleosol carbonate isotopic composition and that of the sample gas.

The sampling method is similar to the flushing method, employing a 2-holed needle and the application of He and As while simultaneously removing headspace. The headspace in this case is a mixture of the equilibrated sample gas and the overpressure gas mixture of He and As. The overpressure gas mixture transports the sample gas from the sample vial through the capillaries into the GasBench II.

At this point a drying stage removes water from the sample gas mixture to maintain a clear passageway through the Valco switch valve and the mass spectrometer inlet valve (Thermo Electron Corp., 2004). The Valco valve cuts a portion of the mixture from the continuous stream by switching to the inject position (Thermo Electron Corp., 2004). The portion is injected into the Gas Chromatograph (GC) column, where a separation in time between CO_2 and other gas components takes place (Thermo Electron Corp., 2004). To decouple the overpressure section of GasBench II from the mass spectrometer's vacuum chamber, the gas mixture passes a second water trap and enters the open split arrangement (Thermo Electron Corp., 2004). While a fixed amount of the gas mixture travels to the mass spectrometer, the excess gas leaves the split to the

surrounding atmosphere (Thermo Electron Corp., 2004). As the original mixture of gases travels to the mass spectrometer source, they are separated by polarity (Thermo Electron Corp., 2004). The travel time for O₂ is approximately 120 seconds, depending on column pressure and temperature. CO₂ requires an additional 20 seconds (Thermo Electron Corp., 2004). The end product is a chromatogram for each sample.

Isodal software was used to run the Gasbench II with acid pump and auto sampler (AS 200-4), the Delta Plus XP Mass Spectrometer, and the TC/EA (Environmental Analyzer). The TC/EA is used to ensure that the peak is centered between each sample, in order to stay focused on the CO₂ off-gassing from the sample.

Subsequent runs of the samples produced elevated CO₂, outside the normal range. Additional measures were taken to account for this problem, including adding additional acid to completely submerge samples with low carbonate content.

Results

Field Results

The measured sections collected at the Conata Picnic area show a clear disconformity between the grey-paleosols and the overlying pink-brown, ledge-forming paleosol. This is particularly apparent in Figure 12. The massive ledge forming paleosol contains a nodular zone approximately one-half meter above the disconformity. This links the ledge forming material clearly with the lower Scenic Member, and Chron C12R; the underlying grey paleosol is therefore the Chadron Formation in Chron C13N. The contact in measured section one is 1.35 meters below ground surface (mbgs), and at measured section two it is 0.8 mbgs.

At Chamberlain Pass the LN zone is approximately 3.1 meters below the top of the measured section. Approximately 1.5 m below the LN zone, the top of the Chadron “haystack” is encountered—a grey paleosol covered with purple chalcedony chips. The Chadron-Brule contact is approximately 4.5 meters below the top of the measured section at Dillon Pass. The Chadron appears to unconformably overlie the paleosol beneath. Just beneath the contact is a white, chalky layer containing chalcedony chips.

Oxygen Isotope Results

Oxygen and carbon isotope measurements from paleosol carbonate are listed in Table 2 and graphical representations are given in Figures 15-18. The measured $\delta^{18}\text{O}$ compositions span a range of -13.99‰ to -5.49‰, with the largest variation occurring in the Conata Picnic Area, measured section 1 (Conata 1). Across all sections variability in the $\delta^{18}\text{O}$ record is notable through the upper Eocene, followed by a strong depletion in the late Eocene from about -6‰ to approximately -10‰ to -14‰. Following the depletion a marked enrichment event occurs just above the Eocene-Oligocene transition, as $\delta^{18}\text{O}$ records reverse trend and tend towards more positive values.

Samples with higher amounts of carbonate were generally run only once, while those with lower amounts were run at least twice, where possible, for better precision. CO_2 yield is reported in Table 1. Errors indicating agreement between duplicate runs of samples are reported as part of Table 2. Sample errors span the range of 0.06‰ to 0.93‰, where 0.2‰ represents a 1-sigma of standard deviation. The majority of errors are near a 1-sigma deviation, but several outliers exist. For samples that were only run once, a 1-sigma error was assumed.

In the Dillon Pass samples, no errors were less than 0.20‰, this section also had the highest reported errors of 0.93‰ and 0.75‰. This is due to very low amounts of CO₂, <0.5 Vs/mg in two samples. For reference, the CO₂ yield of the standards, which are pure carbonates, approximately 75 Vs/mg.

The Chamberlain Pass samples also contain a few confounders. Sample MM-60 at 6.9 mbgs had too little carbonate to allow for more than one run, therefore precision is poor. Unfortunately, it is also the most negative $\delta^{18}\text{O}$ for the section, at -12.10‰ +/- .20‰. The sample was obtained from a white layer that was strongly calcareous on the surface. Fortunately, sample MM-59 was collected and analyzed from closer to the surface of this same layer. That sample has a $\delta^{18}\text{O}$ value of -5.58‰ +/- 0.2‰. This is not in agreement with the buried sample MM-60, being 6.52‰ further depleted in $\delta^{18}\text{O}$. In comparison with other measured sections, the matter is not resolved. In Dillon Pass only one other carbonate sample was collected at a deeper interval, and the $\delta^{18}\text{O}$ of that sample was approximately -6‰. Stratigraphically lower samples from Conata Picnic Area 2 (Conata 2) were similarly in the -6‰ to -7‰. However, Conata 1 shows a value of -11‰ lower in the section. Therefore, the other sections lend support to either $\delta^{18}\text{O}$ value (-5.58‰ or -12.1‰) obtained at 6.9 mbgs at Chamberlain Pass. The value of -12.10‰ is used in this study due to the sampled paleosol (MM-60) being buried and unlikely to be influenced by diagenesis, as compared with the more surficial sample (MM-59). Additionally, Conata 2 does not extend as far below the Eocene-Oligocene boundary as any other sampled section. Therefore, it is possible that had additional samples below three meters been possible to obtain, a more negative $\delta^{18}\text{O}$ may have been obtained, similar to Conata 1, which had a $\delta^{18}\text{O}$ result of -10.64 at 4.30 mbgs. The most scientific

solution would be to throw out both results and re-sample to obtain a true value. The lack of samples above or below in the section means that throwing out that sample would result in approximately 3.5 meters of un-sampled section.

All sections demonstrate enrichment of $\delta^{18}\text{O}$ (trending towards more positive values) shortly after the Eocene-Oligocene transition. At Chamberlain Pass the shift is from -10‰ to -9‰, which is small in comparison to the other sections. At the Conata Picnic area the shift is from -14‰ to -9‰ and -11‰ to -8‰, while at Dillon Pass $\delta^{18}\text{O}$ is enriched from -12‰ to -6‰.

Diagenesis of soil carbonates is an important potential problem. Diagenesis is not suspected in samples collected at the Badlands due to similar trends across the park.

When calcite nodules are partially re-crystallized to sparite, the $\delta^{13}\text{C}$ values are generally about the same, but the $\delta^{18}\text{O}$ values are generally depleted in the sparite by up to 10‰, indicating re-crystallization under higher temperature conditions (Cerling, 1991). The only sample suspected of diagenesis was the surficial sample collected at 6.9 mbgs at Chamberlain Pass (MM-59). This sample was not included for discussion due to likely diagenesis and skewed $\delta^{18}\text{O}$. Trends in the record discussed are therefore primary, rather than a secondary overprint, and reflect syndepositional changes in the $\delta^{18}\text{O}$ value of temperature.

While the errors in the $\delta^{18}\text{O}$ shift are not nominal, they are not large enough to outweigh the overall signal. Across all sections the upper Eocene is marked by wide variation in $\delta^{18}\text{O}$ values, followed by enrichment of $\delta^{18}\text{O}$ after the terrestrial Eocene-Oligocene transition anywhere from approximately 1‰ up to 6‰.

Carbon Isotope Results

Carbon isotope measurements from paleosol carbonate are listed in Table 2, and graphical representations are given in Figures 15-18. The measured $\delta^{13}\text{C}$ compositions span a range of -5.74‰ to -9.97‰, with the largest variation occurring in the Chamberlain Pass measured section. Sample errors span the range of 0.01‰ to 0.46‰, where 0.1‰ represents a 1-sigma of standard deviation. The majority of errors are near a 1-sigma deviation, but several outliers exist.

All measured sections show relative stability in the carbon record through the upper Eocene. Besides the outlier value of -9.97 at MM-60 in Chamberlain Pass, remaining values are tightly clustered around -6.5‰. At Conata 1 the $\delta^{13}\text{C}$ gradually becomes more positive prior to the Eocene-Oligocene, and continues on this trend after the transition. Overall there is approximately a 1.3‰ decrease at this location. At Conata 2, the trend is similar but more pronounced. At the same time that the $\delta^{18}\text{O}$ dramatically gains 3‰, the $\delta^{13}\text{C}$ also gains 0.3‰. At Chamberlain Pass and Dillon Pass the signal varies a bit, but is largely similar. A gradual increase in the $\delta^{13}\text{C}$ is noted from before the Eocene-Oligocene until just after the transition in Chamberlain Pass. Immediately after the transition, $\delta^{13}\text{C}$ rebounds to nearly pre-transition values. Within the overlying 0.2m, $\delta^{13}\text{C}$ shifts back towards more positive values by 0.9‰. This trend continues up to 0.7m from the top of the measured section, with a change in $\delta^{13}\text{C}$ of 0.29‰. At Dillon Pass the $\delta^{13}\text{C}$ signal actually shows strong similarity to the $\delta^{18}\text{O}$, with a marked shift towards more positive values by approximately 1.4‰ coeval with the step in the $\delta^{18}\text{O}$, and the trend towards more positive values continues up-section.

The variations in $\delta^{13}\text{C}$ are similar in many ways to the $\delta^{18}\text{O}$ results. Dillon Pass once again has the samples with the highest errors, 0.22‰ at MM-39 and 0.46‰ at MM-42, corresponding to the same samples with the highest $\delta^{18}\text{O}$ error. This is due to low carbonate content in the sample, yielding poorer precision. Errors in the $\delta^{13}\text{C}$ data may override any interpretation of the trend. Sample 18 has a $\delta^{13}\text{C}$ value of -7.54‰ with an error of 0.46‰. The true value of sample MM-42 could therefore be -7.08‰. Similarly, sample MM-39 has a $\delta^{13}\text{C}$ value of -6.91‰ with an error of 0.22‰. It could therefore have a true $\delta^{13}\text{C}$ value of -7.13 ‰. If both of these are true, then the small shift towards less negative $\delta^{13}\text{C}$ coeval with the shift in $\delta^{18}\text{O}$ could effectively be reduced to a gradual and small shift across the entire measured section.

Chamberlain Pass samples have the lowest errors, ranging from 0.01‰ to 0.18‰. Again, at sample MM-60 too little carbonate was obtained for more than one run and the precision is very poor. An error is not given, so 0.10‰ is assumed. Similar to the $\delta^{18}\text{O}$ value, the $\delta^{13}\text{C}$ value at this interval is the most negative in the section (-9.97‰). Since diagenesis does not alter the $\delta^{13}\text{C}$ value (Cerling, 1991), it may be possible to rely on the $\delta^{13}\text{C}$ value obtained from the surface sample (MM-59), -7.21‰, even though the $\delta^{18}\text{O}$ value is not useable. The value of -7.21‰ from MM-59 is nearly identical to the $\delta^{13}\text{C}$ value in Dillon Pass at the same depth. Some value between the two reported $\delta^{13}\text{C}$ values at 6.9 mbgs in Chamberlain Pass is likely to be the true result. In any case, use of -7.21‰ or -9.97‰ does not change the overall signal, just the magnitude of the $\delta^{13}\text{C}$ shifting towards less negative values at that depth.

CHAPTER VII: DISCUSSION

The marine Oi-1 event, marking the global change in climate from “Greenhouse” to “Icehouse” is well established at the Eocene-Oligocene boundary. While the record is clear that climate deteriorated, the cause is up for debate. The two main camps call for either considerable decrease in atmospheric CO₂ or the buildup of the Antarctic Ice Sheet due to tectonic forcing and subsequent influence of the circum-polar current re-organizing global ocean circulation patterns.

Of further interest, the sediment stratigraphy, fossil record and isotopic studies from the Gulf Coastal Plain, USA indicated that the global climatic deterioration at the Eocene-Oligocene was not ubiquitous (Ivany et al., 2003; Miller et al., 2008; Fluegeman, 2009). The global record may have preceded the near-shore response, and may have had varying effects as local conditions prevailed. The apparent decoupling between the Gulf Coast and the deep marine led to questions concerning the effect of global climate change within the interior continent. The Badlands White River Group has been extensively studied in terms of fossil assemblages and depositional sequences, but a precise temperature chronology was lacking. Zanazzi et al. (2007) completed an isotopic study

on fossil teeth in an attempt to reconcile the mid-continent with the Gulf Coast and global records. Their results, as discussed below, showed the same magnitude of change as the global record, but a lag of approximately 400,000 years. The Badlands fossil tooth isotopic record agrees with the Gulf Coast record only as far as demonstrating a regional lag in response to the global trend. The near-shore record also shows increases in seasonality and aridity, and demonstrates variability throughout the upper Eocene. All qualitative assessments of the Badlands region agree with the Gulf Coast record in terms of increasing aridity.

Can an additional mid-continent isotopic record clarify the effects of global change on the region and elucidate the various forces responsible for mid-continent climate deterioration? And does a better understanding of mid-continent response to global climate change explain the Chadronian-Orellan mammal turnover? This study set out to find an additional mid-continental isotopic record from the Badlands National Park. The isotopes were collected from carbonate paleosols, with the intent of comparing them to the fossil tooth record, the near-shore marine record of the Gulf Coast and the global deep-sea record in order to assess the differences between the regions as global conditions deteriorated. The results from the Badlands paleosols will be discussed, followed by a discussion of the global record and the debate regarding the cause of global climate deterioration. A comparison of the Badlands paleosol isotopic record to each of the above-mentioned regional isotopic studies will then take place, including the implications the findings have on our understanding of global climate change and its effects on the terrestrial realm and faunal distribution of the mid-continent.

The Eocene-Oligocene boundary in the Badlands is placed at the Chron C13R/Chron C13N boundary, which coincides with the marine Oi-1 event. The $\delta^{18}\text{O}$ enrichment in the Badlands stratigraphy occurs after the terrestrial Eocene-Oligocene transition in all sections. Therefore, the terrestrial $\delta^{18}\text{O}$ shift lags the marine Oi-1 event. The Badlands paleosol $\delta^{18}\text{O}$ enrichment ranges from approximately 1‰ up to 6‰. The enrichment follows a variable record through the upper Eocene that ends in a 4‰-8‰ decrease in $\delta^{18}\text{O}$. A variable upper Eocene $\delta^{18}\text{O}$ record is in stark contrast to the deep-marine smooth and gradual trend to more positive $\delta^{18}\text{O}$, indicating that the climatic conditions of the Terrestrial Eocene-Oligocene are not straightforward. This data indicates that global decreases in atmospheric CO_2 may not have been the most important factor for climate in the mid-continent. Furthermore, the lag in continental climate change behind the global record presents a decoupling between the terrestrial and deep-marine climatic events.

The $\delta^{13}\text{C}$ record obtained from the Eocene-Oligocene soil carbonates is a potential indicator of atmospheric $\text{P}(\text{CO}_2)$ levels, water stress, plant productivity and canopy structure, all increasing with increasing aridity (Cerling et al., 1991; Zanzzi et al., 2007). Very little change is noted in the $\delta^{13}\text{C}$ record throughout the Eocene-Oligocene sections sampled. At the Conata Picnic Area-1, a gradual but noticeable $\delta^{13}\text{C}$ enrichment of 1.3‰ is noted across the Eocene-Oligocene. The Conata Picnic Area-2 record also shows an enrichment in $\delta^{13}\text{C}$ across the Eocene-Oligocene, though the amplitude is less. At Chamberlain Pass and Dillon Pass, slight $\delta^{13}\text{C}$ enrichment is coeval with the $\delta^{18}\text{O}$ enrichment, occurring after the Eocene-Oligocene. The slight enrichments in the $\delta^{13}\text{C}$ record indicate that at least some water stress or increase in aridity was occurring in the region, although it appears small. Dramatic increases in $\delta^{18}\text{O}$ and subtle, coeval increases

in $\delta^{13}\text{C}$ indicate that temperature was declining rapidly while aridity was likely increasing, although to a much smaller degree.

The Badlands were likely affected by several climate factors: decreasing atmospheric CO_2 and lowering of temperatures, intermittent volcanic activity overprinting the local atmosphere with variable levels of CO_2 , and disrupting atmospheric patterns, glacial development in the Arctic and nearby Rocky Mountains possibly leading to variable distributions of colder temperatures and less precipitation, and a decoupling from the oceanic system leading to an increased magnitude of temperature change.

As discussed throughout this paper, decreasing temperatures and increasing aridity were widespread across the globe at the Eocene-Oligocene. The most cited culprit for global climate deterioration is decreasing atmospheric CO_2 leading to the build up of the Antarctic Ice sheet, but this argument faces much debate. Increased tectonic activity throughout the Eocene, particularly the uplift of the Tibetan plateau, is particularly effective at re-distributing the global CO_2 budget from the atmosphere to oceanic sediments. One argument in favor of gradual atmospheric CO_2 reduction is the $\delta^{18}\text{O}$ record itself. As mentioned previously, the deep-ocean $\delta^{18}\text{O}$ record shows a gradual enrichment indicating colder temperatures throughout the Eocene, followed by an acceleration of climatic deterioration at the Eocene-Oligocene boundary (Figure 5). By itself, this indicates that a global mechanism was acting to slowly decrease temperature until a threshold value was reached at the Eocene-Oligocene, at which point temperature rapidly decreased. Cold deep ocean temperatures at locations of deep-water formation indicate sea-surface temperatures at those regions are also cold, as they must be to initiate sinking. According to Dwyer, 2000, colder sea-surface temperatures indicate the air-sea

exchange of CO₂. Cooler deep-water temperatures may therefore indirectly signify enhanced oceanic CO₂ storage, and concomitant decreases in atmospheric CO₂ (Dwyer, 2000).

Although there is good evidence to support decreases in atmospheric CO₂, at the Eocene-Oligocene transition the level of atmospheric CO₂ was still significantly higher than pre-industrial levels (DeConto and Pollard, 2003). Another argument postulates that development of the East Antarctic Ice Sheet (EAIS) grew with no appreciable drop in temperature (Lear et al., 2000). Growth of the EAIS is attributed instead to the tectonic separation of Antarctica from other southern hemisphere continents, opening of Southern Ocean gateways and subsequently creating the circum-Antarctic oceanic current. While the Tasmanian Passage (between Antarctica and Australia) may have been open at the Eocene-Oligocene, the Drake Passage (between South America and Antarctica) may not have been significantly deep enough to affect climate change until several million years after the Eocene-Oligocene transition (DeConto and Pollard, 2003).

DeConto and Pollard (2003) tested the importance of changing atmospheric CO₂, orbital parameters, and ocean heat transport in the nucleation and fluctuations of the early EAIS using a coupled global climate and ice-sheet model. Their results showed that when both passages were open ice sheet expansion did not occur until atmospheric CO₂ decreased to 3 times pre-industrial atmospheric levels (PAL), with the EAIS established at 2.8 PAL. When the model was run with the Drake Passage remaining closed, heat transport was increased by 20%, requiring atmospheric CO₂ to decrease an additional 140ppmv for major glaciation to occur, but the overall range of atmospheric CO₂ remained the same (DeConto and Pollard, 2003).

The model results show that the tectonic opening of the Drake Passage, and subsequent formation of Circum-Antarctic ocean currents, may not have been as important in forming the EAIS as atmospheric levels of CO₂, but may have ensured its endurance since the Oligocene (DeConto and Pollard, 2003). The conclusion ultimately supports the first global climate change causal argument, that decreases in atmospheric CO₂, particularly past an unsustainable threshold, coupled with orbital forcing and ice-climate feedbacks were the primary causes for the sudden build-up and variations of Antarctic ice at the Eocene-Oligocene boundary (DeConto and Pollard, 2003). But if atmospheric CO₂ levels were the primary driver of ice sheet build-up, why did it only occur on Antarctica?

Recent studies have found evidence of glaciation on Greenland beginning in the Eocene, but not to the extent that it developed on Antarctica (Eldrett et al., 2007; DeConto et al., 2008; Tripathi et al., 2008). Northern Hemisphere ice sheets are not supported until the Miocene (Zachos et al., 2001). The fact that both poles did develop glaciers further supports a decrease in atmospheric CO₂. However, stable ice sheet development in the Northern Hemisphere requires atmospheric CO₂ levels much lower than have been reported for the Eocene-Oligocene (DeConto et al., 2008). The isolation of Antarctica and development of the circum-Antarctic current sets it apart from Greenland, giving Antarctica a favorable advantage for ice sheet development once atmospheric CO₂ decreased below a restrictive threshold.

The deep-marine record, while powerful in scope and indications of global change, leaves much to be desired in terms of resolving the cause of the cooling event. Can the terrestrial and near-shore marine records help to decipher the issue?

The isotopic results from the Badlands National Park presented in this study provide interesting results that argue for some combination of the influence of decreasing atmospheric CO₂, increasing ice-volume with subsequent feedback mechanisms and local tectonic control in controlling the terrestrial shift from “Greenhouse” to “Icehouse”. As previously mentioned, the isotopic records obtained by Zanazzi et al (2007), support this study’s findings that the terrestrial $\delta^{18}\text{O}$ enrichment occurs after the Eocene-Oligocene transition, lagging the marine Oi-1 event. They further find that the continental transition is of higher magnitude than the marine transition (Zanazzi et al., 2007).

Their investigation of fossil tooth and bone isotopic composition found a large drop in Mean Annual Temperature (MAT) of 8.2 +/- 3.1°C over 400,000 years, with small increases seasonality and virtually no change in aridity (Zanazzi et al., 2007). Tooth enamel contains zonations indicating seasonality. The zones analyzed in four mammals had almost indistinguishable $\delta^{18}\text{O}$ records, while bone $\delta^{18}\text{O}$ showed a 1.7‰ positive shift across the boundary (Zanazzi et al., 2007). The difference between the two record sets indicates that between seasons water composition remained nearly unchanged, hence there was no major change in temperature seasonality; while at the same time the bone isotopic composition indicates a decrease in MAT (Zanazzi et al., 2007). Small increases in the range of *Meshippus*, *Merycoiodon*, and *Subhydracodon* enamel $\delta^{18}\text{O}$ may reflect a slightly higher Oligocene versus Eocene mean annual range of temperature (MART) (Zanazzi et al., 2007). Marine records support at most a 2-3°C cooling at low to mid-latitudes, while Zanazzi et al’s MAT difference is approximately 8°C (2007). This difference remains unresolved.

According to Zanazzi et. al. (2007) the lack of seasonality increases, no resolvable change in aridity, and a large scale MAT drop support climatic deterioration due to a major decrease in the atmospheric CO₂. They claim that the opening of the Southern Ocean gateways must have had a minor role in global cooling because the Antarctic Circumpolar Current is expected to produce warming of the middle and high latitudes of the Northern Hemisphere, directly in contrast to their results (Zanazzi et al., 2007). Mid-latitude warming in response to the circum-Antarctic current is supported by the model developed by Sloan and Barron (1992). If the argument presented by DeConto and Pollard (2003) is correct, and the circum-Antarctic current was not formed until after the Eocene-Oligocene transition, then this argument by Zanazzi et al. (2007) is relevant. However, if the circum-Antarctic current was established at the Eocene-Oligocene, then because this action is expected to warm mid and high latitudes in the Northern Hemisphere, it could be responsible for continental temperatures in North America lagging behind the marine ones. It is also likely that the tectonic regime influencing local atmospheric patterns affected the northern Great Plains to a greater degree than previously estimated. Had climate change only been the result of decreasing atmospheric CO₂, the variability seen in the upper Eocene paleosol record would not be expected. In fact, global climate models conclude that local elevation changes play an active role in shaping interior continental climate (Sloan and Barron, 1992).

It is difficult to draw many conclusions from the $\delta^{13}\text{C}$ record obtained from paleosol carbonates. In the Badlands paleosols, it does appear that a slight shift in $\delta^{13}\text{C}$ occurs towards more positive values in association with the large jump in $\delta^{18}\text{O}$, which supports increased aridity. However, the shift is slight (between 0.1‰ and 0.6‰) and not

apparent at Conata measured section-one. This would appear to support the findings by Zanazzi et al (2007) that aridity did not appreciably increase from the Eocene to the Oligocene. A slight shift in the paleosol record, however, is still indicative of increasing aridity in the region. Increasing continental aridity would be expected as global temperatures cool and evaporative transport is lessened—particularly by the rain-shadow effect of the Rocky Mountains and the Black Hills. Furthermore, qualitative records across the continent support the argument for increased aridity and water stress.

In the Badlands National Park, the paleosol sequences have been used to assess the paleoecology from the Eocene to the Oligocene. Forests are indicated in basal paleosols of the Chadron Formation by large root traces and silicified wood with seasonal growth rings (Retallack, 1992). Late Eocene Chadron paleosols also contain large root traces with size and distribution profiles indicating dry woodland along stream banks, and some thin paleosol layers with fine root traces that indicate herbaceous vegetation (Retallack, 1992). Sedimentary evidence indicates stream and flood plain deposits, including shallow warm lakes, and a forested floodplain with scattered sedge meadows, oxbow ponds and forested upland environments (Stoffer, 2003). Gleska Series paleosols in the Chadron have progressively better-differentiated, calcareous horizons than Interior Series paleosols indicating a decline of mean annual rainfall towards the late Eocene (Retallack, 1983).

The presence of Lacustrine Limestones in the Chadron Formation also indicates formation in a humid temperate climate similar to many temperate groundwater-fed, hard water lakes (Evans and Welzenbach, 2000). The paleontology of the Lacustrine Limestone facies includes fish, turtle plates, gastropods, bivalve shell fragments,

ostracodes, calcified aquatic roots, charophytes and algal filaments (Evans and Welzenbach, 2000). There was high abundance and low diversity in the fossils, with evidence of environmental stress—including fish-kill horizons and death horizons of ostracodes (Evans and Welzenbach, 2000).

The woodlands of the Eocene gave way to savannah in the Oligocene (Stoffer, 2003). The Scenic Member of the Brule Formation displays large drab haloed root traces, hackberry endocarps, early herbaceous plants in streamside swales, and scattered trees with intervening areas of bunch grasses and small shrubs in the interfluves (Retallack, 1983). The early Oligocene in the Badlands region can be described as well within the sub-humid range (Retallack, 1983). Only fine root traces are seen in the A-C horizons of former high terraces and flood plains (Pinnacles and Samna series soils) (Retallack, 1983). This, along with cracks and veins in the paleosols indicate a vegetative cover of scattered bunch grasses and shrubs in the Pinnacles series (Retallack, 1983).

Streams migrated across the savannah-like floodplain due to low relief (Stoffer, 2003). Volcanic ash and alluvium were deeply weathered under the wooded landscape of confined, meandering alluvial channels (Retallack, 1983). Weathering cannot be due to tectonic changes alone because changes in the paleotopography and paleochannels are not striking (Retallack, 1983). Changes in weathering are most likely the result of decreasing stability of the landscape due to decreasing vegetative cover and increasingly dry climate (Retallack, 1983).

Mean Annual Rainfall (MAR) can be estimated using Jenny's (1941, 1980) relationship between depth to calcic nodules and rainfall. The depth of the calcic horizons in the Badlands paleosols compared with the depths in soils of the present Great Plains

indicates that the late Eocene (38 Ma) mean annual rainfall was greater than 1000 mm, but by 32 Ma, during Oligocene Scenic Member deposition, it was down to 500-900 mm (Retallack, 1992). Based on the petrocalcic horizons MAR continued to decline throughout the Oligocene (Retallack, 1992). Petrocalcic horizons and chalcedony pseudomorphs of Barite on Gypsum roses in some Gleska series paleosols indicate severe droughts during the late Oligocene (Retallack, 1983). A lack of tufa into the Oligocene is a further indication of increased aridity, as decreasing precipitation could have affected regional recharge-discharge relationships, water table positions, and rates of carbonate precipitation (Evans and Welzenbach, 2000).

Clay types in paleosols also indicate declines in MAR. Lower paleosols in Badlands sections contain greater amounts of kaolinite, which is found in humid environments (Retallack, 1992). Stratigraphically higher paleosols contain greater amounts of smectite and increasing amounts of illite, indicating drier climate (Retallack, 1992). Evidence for very dry and highly seasonal climates was localized to boundaries between rock units; for example, the Chadron immediately below the Brule Formation is riddled with chalcedony replaced gypsum pseudomorphs and chalcedony-filled deep cracks (Retallack, 1992). This type of mineralization has been documented in saline playa lakes and surrounding soils and suggests climate was even drier at times than precipitation estimates suggest (Retallack, 1992).

It is difficult to ignore this evidence in favor of confounding $\delta^{13}\text{C}$ records from paleosol carbonates. Therefore, it is likely that the $\delta^{13}\text{C}$ records from Badlands paleosols are indicators of increasing plant stress or aridity, and in disagreement with the conclusions drawn by Zanzazi et al. (2007). Given that Holocene records of the mid-

continent demonstrate that even small changes in precipitation had a profound effect on geomorphic responses (Denniston et al., 2007), the small change in Badlands $\delta^{13}\text{C}$ across the transition may have been enough to cause the dramatic record of increased aridity in paleosol development and fossil assemblages.

The disagreement between the two isotope studies conducted on deposits and fossils within the White River Group and the paleoecological assessments made with more subjective criteria requires further investigation. Perhaps our understanding of the role aridity plays on isotopic distributions is not well understood.

The isotopic composition of soil carbonate is also a proxy for atmospheric CO_2 . In general, decreasing $\delta^{13}\text{C}$ of marine carbonates indicates increasing atmospheric CO_2 (Zachos et al., 2001). Slight increases recorded in the $\delta^{13}\text{C}$ from three of the measured sections after the Eocene-Oligocene boundary indicate that atmospheric CO_2 did in fact decrease. However, the relative stability in the record indicates that atmospheric CO_2 did not abruptly change across the Eocene-Oligocene in the mid-continent. This is evidence against the first argument for global climate deterioration, which postulates that global decreases in atmospheric CO_2 were the cause. This is also in contrast to the Zanazzi et al. (2007) conclusion that a dramatic decrease in atmospheric CO_2 must have been the cause of mid-continent climate deterioration.

Magnetostratigraphy of the region has been useful in linking Badlands stratigraphy to disparate sections, particularly in comparing the local stable isotope record to the global record. Paleomagnetic stratigraphy offers the possibility of globally synchronous time correlation and is the only method that allows high-resolution correlation of the land sequences with the marine record (Prothero and Swisher, 1992).

The “lower nodular zone” (LN zone) used by most collectors to zone their specimens was used to reference sections in the field to existing magnetostratigraphic studies, as was discussed in the Field Results section of Chapter VI (Figure 4).

At the Chamberlain Pass Scenic type section, the LN zone marks the division between the Chadron Formation and the Scenic Member of the Brule Formation, and occurs within Chron C13N (Prothero and Whittlesey, 1998). At the Pig Dig region of the park, however, an unconformity separates the Chadron Formation from the Scenic Member, which is overlain by the LN zone (Prothero and Whittlesey, 1998). This was observed in the field and is depicted in Figure 12. The LN zone and the uppermost portion of the Chadron Formation lie within Chron C13N (Prothero and Whittlesey, 1998). An unconformity also separates the Chadron Formation from the Scenic member in the Pinnacles, but the base of the Scenic remains in the LN zone (Retallack, 1983). Observations in the field at Dillon Pass confirm this. According to Prothero the LN zone lies within C13R, as does the uppermost portion of the Chadron formation at the Pinnacles location (1992). Chron C13N seems to have been removed by the unconformity at this measured section, so the bulk of the Chadron Formation is likely in C13N, but is not directly measured (Prothero and Whittlesey, 1998).

To summarize, the section at Chamberlain Pass (Scenic type section) lies within C13N at the LN zone, the contact between the Chadron and the Brule. At the Pig Dig area, the Conata Picnic Area measured sections show the contact to lie within Chron C13N and is overlain by the LN zone. At the Pinnacles area, the contact is within Chron C13R (Figure 4).

The magnetostratigraphy of the Badlands measured sections also allows for direct comparison of the temperate mid-continent isotopic record to that of the sub-tropical shallow marine record in the Gulf Coast, USA. The Eocene-Oligocene transition in the Gulf Coast takes place within Chron C13R, with the Oi-1 event occurring in the early Oligocene at the transition from Chron C13R to Chron C13N (Miller et al., 2008).

The Gulf Coast presents complicated stratigraphy across the Eocene-Oligocene boundary. From the different lines of evidence provided from several studies of the Gulf Coastal Plain, it appears that the Oi-1 event had a significant influence on the stratigraphy and biota of the region. As previously mentioned, the unconformity at the Yazoo-Red Bluff contact was produced by a fall in relative sea level possibly associated with Oi-1 (Fluegeman, 2009; Fluegeman et al., 2009). Additionally, the turnover in Jackson-Vicksburg benthic foraminifera occurs after the Eocene-Oligocene boundary in the Gulf Coast, making it a likely effect of the marine Oi-1 event. Further evidence for an Oi-1 influence on the region is discovered when analyzing the paleoecological assemblages of the region. The Yazoo Clay contains benthic foraminiferal assemblages indicating outer-neritic to upper-bathyal conditions, while the Red Bluff assemblage indicates middle-neritic (Fluegeman, 2009). This is an indication that the Eocene-Oligocene boundary in the Yazoo Clay occurs not during low sea level, but actually at maximum water depth (Fluegeman, 2009).

Sudden shallowing is indicated by the unconformity and also by the planktic:benthic (P:B) ratio. P:B ratios are indicative of sea-level fluctuations, with higher ratios indicating deeper waters. In all three cores of the Chickasawhay Valley, the P:B ratios are relatively high but fall abruptly in the lower Red Bluff Clay, suggesting a

regressive shoreline shift at the Yazoo-Red Bluff contact (Fluegeman, 2009). Moving up through the lower part of the Red Bluff Clay, the P:B ratio increases, indicating a landward shift of the shoreline (Fluegeman, 2009). This is likely the first postglacial transgression in the eastern Gulf Coastal Plain (Fluegeman, 2009). These findings correspond nicely to those found in the Badlands paleosol isotopic record. Decoupling of the near-shore marine and terrestrial records is evident as the terrestrial Oi-1 event appears to lag the Eocene-Oligocene in the Gulf Coast as well.

Some confusion does come into play when comparing to other nearby sections though. At Little Stave Creek, Clarke County, Alabama, the foram assemblages from the Yazoo to Red Bluff are the same as at other sections; however, the distribution of Jackson-Vicksburg taxa across the contact is different, with many species having their highest and lowest occurrences not at the contact, but at different levels near the contact (Fluegeman, 2009). This same pattern is found at the closely related St. Stephens Quarry, and at the Mossy Grove Core in Mississippi (Fluegeman, 2003, 2009). This data contradicts the earlier conclusion that the Yazoo-Red Bluff contact is a sequence boundary. A complex stratigraphic relationship between the Yazoo-Red Bluff sections of the Chickasawhay Valley of Mississippi and those of southwestern Alabama can be inferred (Fluegeman, 2009). It is useful to note that similar patterns in the biostratigraphy as well as a lithologic unconformity are noted in northwestern Europe and Australia at this time period (Fluegeman, 2009).

The Gulf Coast $\delta^{18}\text{O}$ record is similar, but not identical to the deep-marine record. Through the upper Eocene there is a trend towards more positive $\delta^{18}\text{O}$, but it is more variable than the global trend (Miller et al., 2008) (Figure 19). In the Gulf Coast the $\delta^{18}\text{O}$

enrichment is marked by a nearly 0.5‰ increase in the upper Eocene (Miller et al., 2008); in the Badlands paleosols the shift is much larger, anywhere from 1‰-6‰. The Gulf Coast $\delta^{13}\text{C}$ record also differs from that of the deep marine. The global record shows a total enrichment across the Eocene-Oligocene of approximately 0.5‰ (Zachos et al., 2001), while the Gulf Coast record indicates enrichment just over 1‰ was stronger than the change in $\delta^{18}\text{O}$ (Miller et al., 2008). By contrast, the magnitude of the $\delta^{13}\text{C}$ enrichment from the Badlands paleosol record is between 0.1‰ and 0.6‰. These measurements are inconsistent with the Gulf Coast records, but do agree in overall trend. Further understanding of the interplay between the mid-continental sections of the Badlands National Park and the Gulf Coast as climate cooled across the region may be gained from comparison to a higher resolution record from the coastal plain.

Otoliths, or inner ear bones, of fish in the Gulf Coast were examined as an independent line of evidence from sedimentation to assess the role of changing global climate on the Gulf Coast (Ivany et al., 2000). Otoliths continue to accrete new material over the life of a fish (2-3 years), allowing for seasonal responses to climate to be recorded (Ivany et al., 2000). Mean $\delta^{18}\text{O}$ values of otoliths increased only slightly across the Eocene-Oligocene boundary, and when mean paleotemperatures are calculated, the differences in temperature across the boundary were not found to be statistically significant (Ivany et al., 2000). However, when analyzed for seasonal effects, the results are much more telling. Late Eocene temperature ranges were from 15°C to 20°C, with minimum winter temperatures of 13.5°C (Ivany et al., 2000). The Oligocene record contains similarly warm summer temperatures, around 20°C, but winter temperatures consistently fall below those of the Eocene—dipping as low as 11°C (Ivany et al., 2000).

Summer temperature differences were not found to be statistically significant, but winter temperatures in the Oligocene were colder than those in the Eocene by as much as 4°C or more (Ivany et al., 2000). A comparable drop in temperature was seen in the montane floras of the Rocky Mountains from 11-6°C (Hutchison, 1982). This Gulf Coast record supports the findings of the Badlands paleosols $\delta^{18}\text{O}$ more so than the record obtained by Miller et al. (2008). Though the Badlands paleosol isotopes do not record seasonality, wide fluctuations are noted throughout the Eocene.

The variability in the regional Gulf Coast carbonate isotopic record, combined with the otolith record demonstrates that a decrease in atmospheric CO_2 cannot be the only factor influencing climate change in the Gulf Coast. Local variances must be linked to changes in oceanographic and/or atmospheric circulation. The fact that a $\delta^{18}\text{O}$ shift is apparent in the Gulf Coast— though to a lesser degree than the deep sea record—and mean annual temperature underwent minimal change across the boundary while winter temperatures became more severe, supports the likelihood that global ice volume was an important factor for the Gulf Coast, not temperature change alone (Ivany et al., 2000). The role of ice volume in the mid-continent may have come in the form of montane glaciers forming in the Rocky Mountains, and ultimately affecting atmospheric circulation, temperature and precipitation patterns.

Further support for the role of ice volume on climate at the Gulf Coast comes from the Moody's Branch Formation and Lower Yazoo Clay from the Mossy Grove Core (Hurley and Fluegeman, 2003). P:B ratios, as well as $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of forams were analyzed to assess the temperature and sea-level changes during the late-middle Eocene (Hurley and Fluegeman, 2003). An inverse relationship, where minimum P:B ratios

coincided with maximum $\delta^{18}\text{O}$ values, and vice versa (ie- shallow water coincided with colder temperatures) was noted (Hurley and Fluegeman, 2003).

It appears that Glacioeustacy may have been a significant controlling factor of sea-level fluctuations by the Bartonian (Hurley and Fluegeman, 2003). This finding, coupled with that of Ivany et al. (2000), lends further support to the arguments that ice was accumulating on Antarctica during the late Eocene, and the dramatic shift in global climate occurred after some threshold value was reached at the Eocene-Oligocene transition. This may also support the argument by Lear et al. (2000) wherein they claim that a drop in atmospheric CO_2 was more important in forming the EAIS than temperature decreases. The observation that a uniform decrease in atmospheric CO_2 is not the primary controlling factor on Gulf Coast climate (Ivany et al., 2000; Hurley and Fluegeman, 2003), and that the Gulf Coast response lagged behind the deep-sea Oi-1 event (Ivany et al., 2000; Fluegeman, 2009; Fluegeman et al., 2009) supports the idea that although atmospheric CO_2 appears to have gradually decreased past an important threshold at the Eocene-Oligocene allowing for the establishment of the EAIS from existing Antarctic glaciers, there was not a concomitant acceleration in temperature deterioration or further dramatic decline of atmospheric CO_2 into the Oligocene. Once the EAIS was well established, and oceanic circulation patterns had reorganized, in part due to changing tectonic regimes, the effects truly became global as they were translated to near-shore marine environments and eventually inland to the mid-continent.

The isotopic records obtained from the Badlands are in agreement with the Gulf Coast climate record to a large degree, with some important differences. Both Badlands isotopic studies found a decoupling between the terrestrial record and the marine record,

which is similarly expressed in the Gulf Coast. However, the isotopic analysis conducted by Zanazzi et al. (2007) indicates a large drop in Mean Annual Temperature with no associated increase in aridity. The temperature findings of Zanazzi et al. (2007) are supported by the Badlands paleosol isotopic record, which demonstrates accelerated enrichment of $\delta^{18}\text{O}$ after the Eocene-Oligocene boundary. Despite the relative stability in the paleosol $\delta^{13}\text{C}$ record, increasing aridity is noted both across the Eocene-Oligocene transition, and in conjunction with the $\delta^{18}\text{O}$ enrichment. This is supported by the qualitative records of the Badlands. Increasing aridity is also found in the Gulf Coast, associated with increased seasonality (Ivany et al., 2000). (Figure 19).

Though the Gulf Coast record shows more variability than the deep marine record, it is more stable than the Badlands paleosol record. Less variability is expected in Gulf Coast sections due to their position at lower latitudes and the fact that changes occurred more dramatically at higher latitudes (Ivany et al., 2000). Zanazzi et al. (2007), claim that their findings support studies that attribute the climate cooling transition to a significant drop in atmospheric CO_2 . Their results do support increased seasonality, but changes are small, and only reflected in temperature, not aridity (Zanazzi et al., 2007); whereas the fish otolith isotopic analysis showed sharp decreases in winter temperatures, and variability that could not be explained by decreasing atmospheric CO_2 alone (Ivany et al., 2000).

The Badlands paleosol isotopic record appears to bridge the gap between the Zanazzi terrestrial isotopic record arguing for decreased atmospheric CO_2 and the Gulf Coast records indicating no dramatic temperature drop and the importance of ice volume over atmospheric CO_2 . The degree of isotopic shift in Badlands paleosols agrees with the

dramatic decrease in temperature found in the fossil bone record of Zanazzi et al. (2007); while the variability in the paleosol record through the upper Eocene corresponds to variability in the Gulf Coast record (Miller et al., 2008), indicating that the climatic conditions of the Terrestrial Eocene-Oligocene are not as straight forward as either Ivany et al. (2003) or Zanazzi et al. (2007) hypothesize.

It is likely that variability is due to altered atmospheric and oceanic circulation patterns, or even to the influence of ice rather than decreasing atmospheric CO₂. Given that the variability in the terrestrial record is larger than that in the marine record, local atmospheric patterns must have had a large influence on mid-continental climate change. It is widely established that climate across the Midwest region of the United States has been strongly influenced by three air masses since the Holocene—the warm and humid Gulf of Mexico, the cooler and drier Pacific, and the cold and dry Arctic (Denniston et al., 2007). Throughout the ice-free Eocene it is reasonable to assume that the differences between the three air masses were much less pronounced than today. Particularly the Pacific and Gulf of Mexico air temperatures could not have been drastically different, given the congruity between faunal and floral assemblages between the two regions, as previously discussed.

As the northern high latitudes cooled, the Pacific and Arctic air masses would be expected to carry colder and drier air across the upper Midwest. Although Northern Hemisphere ice sheets are not supported until the late Miocene, evidence of ice formation on Greenland has been found (Eldrett et al., 2007; Tripathi et al., 2008), lending support to the possibility for small montane glaciers to have formed in the nearby Rocky Mountains given their high altitude and particularly high latitude of the northern reach of the Rocky

Mountains. Ice growth is expected to be strongest in the winter, during which time the Arctic and to a lesser degree, Pacific, air masses would have become markedly colder and drier, likely losing precipitation to the glaciers. These conditions would then be reflected in the precipitation they brought with them to South Dakota and the Midwest at large.

Indeed, increased seasonality is predicted by changes in local elevation (Sloan and Barron, 1992), and noted in the isotopic records of both Zanazzi et al. (2007) and Ivany et al. (2003). Had climate change only been the result of decreasing atmospheric CO₂, uniform change would be found in all regions (Ivany et al., 2000). Furthermore, although the terrestrial and the Gulf Coast records indicate decoupling from the deep-marine record, further decoupling of the Badlands from the Gulf Coast is supported by the large shift in Badlands $\delta^{18}\text{O}$. It is likely that the local climatic regime and controlling factors play a larger role in terrestrial climate than in the near-shore marine environment.

A comparison with Asian and European floral and faunal evolutionary patterns at similar latitudes to the Badlands indicates similar changes in vegetation across the Northern Hemisphere at the Eocene-Oligocene boundary. Decreasing temperature and increasing aridity are associated with the disappearance of most of the tropical species, which had been very widespread during the late Cretaceous and Eocene (Cavelier et al., 1981). The inland Paris Basin floral community indicates a marked change in the late Eocene following the gradual diminution of tropical species to more temperate species through the Eocene (Cavelier et al., 1981).

During the early Eocene, Northern Hemisphere reptiles lived at latitudes poleward of 70° (Sloan and Barron, 1992). A mass immigration of North American land mammals and several lizard families across the Thulian land bridge into Europe is noted in the early

Eocene (Briggs, 1987). In northwest China, subtropical woody savanna developed during the early Eocene, lacking evidence of grasses and herbaceous pollen until the late Eocene at which time sub-arid floras are found (Prothero, 1994).

The Grande Coupure of European large mammals has also been observed in rodents and reptiles and various geographic locations (Cavelier et al., 1981). After the many sizes of mammals that existed in the warm Eocene of Asia, small rodents and lagomorphs dominated the Asian Oligocene, as they were better adapted to the arid and cool environment of the Oligocene (Hartenberger, 1998; Meng & McKenna, 1998). Similar changes were noted in Europe, allowing for the conclusion that the restructuring of mammalian communities in Europe was not only a consequence of Asian invasion, but overall global climate trends (Hartenberger, 1998). Indeed, similarities across the Northern Hemisphere are likely due to a general drop of temperature accompanied by a period of great drought (Cavelier et al., 1981).

Large herbivores suffered at the Eocene-Oligocene due to new, un-diversified immigrants of Asian origin replacing most of the old endemic Eocene genera of Europe (Cavelier et al., 1981; Hartenberger, 1998). Extinctions and low diversity of the new faunal immigrants appear to be the result of climatic deterioration (Cavelier et al., 1981). Asian immigrants were also thought to have influenced the reworking of North American faunas, allowing many Asian families and genera to reach cosmopolitan status (Hartenberger, 1998).

A large change in mid-continent MAT exceeding changes in sea surface temperatures at comparable latitudes explains why faunal turnover is noted for gastropods, amphibians and reptiles, but mammals were unaffected (Zanazzi et al., 2007).

Pulses of temperature and aridity stresses throughout the upper Eocene, as evidenced in the variability of the isotopic record from the Badlands paleosols, likely created significant selection pressure in the region. This coupled with significant influx of immigrants may have led to a White River mammal regime that was better adapted to changing conditions. An 8°C drop in temperature could have caused Oligocene temps to approach freezing, explaining the demise of many terrestrial heterotherms while most large mammals survived (Zanazzi et al., 2007). The Chadronian-Orellan NALMA may be due in part to climatic change in the region, and in part due to immigration of competing species. The fact that the Oi-1 event is not expressed in the mid-continent until after the Chadronian-Orellan NALMA argues against a significant climate change event causing faunal extinctions.

In light of the findings of the Badlands paleosol isotopes determining that a large temperature drop was found after the Eocene-Oligocene with an associated small decrease in aridity that led to large changes in faunal and floral assemblages, the similarity of plant and animal fossil trends across the Northern Hemisphere may indicate that climate change across the terrestrial realm was expressed in a similar manner.

CHAPTER VIII: CONCLUSIONS

During establishment of the EAIS, and subsequent surficial and deep-water re-organization of the global heat budgets the near-shore and mid-continent climates remained relatively stable. Ice volume influence is noticeable, particularly in the Gulf Coast, but temperature and humidity did not immediately deteriorate. If a critical threshold in atmospheric CO₂ was breached at the Eocene-Oligocene transition, causing the development of the EAIS and establishing the Oi-1 marine record, it was not noted in the mid-continent Badlands paleosol sequences. Variability in the upper Eocene $\delta^{18}\text{O}$ from Badlands paleosols indicates that local tectonic arrangements and climatic regimes were the likely forces keeping the global transition to “Icehouse” at bay.

After the marine Oi-1 event occurred and feedback mechanisms gained strength, a critical threshold was reached in the mid-continent. It appears that once this final threshold was attained, there were no remaining buffers in the northern Great Plains to deflect the changing global climatic regime. Prevailing global changes gained dominance as local regimes gave way, and temperatures in the mid-continent decreased significantly. This is clearly preserved in the $\delta^{18}\text{O}$ record obtained from paleosols within the Badlands

National Park. This appears to parallel the changes that occurred on Antarctica that led to the initial build up of the EAIS.

The terrestrial lag behind the marine Oi-1 event is not unexpected. The Oi-1 marine event is recorded in deep-sea benthic forams. Therefore, a rapid reduction in deep-sea temperatures related to the formation of continental ice on Antarctica and subsequent deep-water reorganization would not be expected to create a coeval climatic impact in the shallow marine or terrestrial realms.

Variability in the upper Eocene $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records from the Badlands paleosols demonstrates pulses of temperature and aridity stresses in the region that would have caused significant selection pressure leading up to the ultimate decline in temperature. This likely explains why cold-blooded fauna experienced more significant turnover than mammals; particularly when coupled with significant influx of mammal immigrants well suited to a Savanna type diet. Given that the marine Oi-1 event is not expressed in the mid-continent until after the Chadronian-Orellan NALMA, it is unlikely that a significant climate change event caused the faunal extinctions.

It is clear that further study of mid-continent sections is warranted. Several depositional basins are well researched across the Eocene-Oligocene transition that would benefit from a quantitative examination of climate change. The two existing isotopic studies of the Badlands are in agreement regarding temperature change, but are not resolved in terms of aridity. Furthermore, the similarity of faunal and floral fossil trends across the Northern Hemisphere may indicate that climate change within the terrestrial realm had a similar effect, and in all cases may have been expressed differently than in the marine record.

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BIBLIOGRAPHY

Barrett, Peter, 2003. Cooling a Continent: *Nature*, Vol. 421, pp 221-223.

Bates, Robert Latimer, Jackson, Julia A., 1984. *Dictionary of Geological Terms*, Third Edition: American Geological Institute, New York.

Bestland, Erick A., Retallack, Gregory J., Swisher, Carl C. III, 1997, Stepwise Climate Change Recorded in Eocene-Oligocene Paleosol Sequences from Central Oregon: *The Journal of Geology*, v. 105, p 153-172.

Birkeland, Peter W., 1974. *Pedology, Weathering and Geomorphological Research*; Oxford University Press, New York.

Blakely, Ron, 2009. Department of Geology, Box 4099 Northern Arizona University, Flagstaff, AZ. <http://jan.ucc.nau.edu/~rcb7/index.html>.

Bottomley, Richard; Grieve, Richard; York, Derek; & Victor Masaitis, 1997. The Age of the Popigai Impact Event and its Relation to Events at the Eocene/Oligocene Boundary: *Nature*, Vol 388, pp 365-368.

Bowen, Gabriel J., Koch, Paul L., Gingerich, Philip D., Norris, Richard D., Bains, Santo, Corfield, Richard M., 2001. Refined Isotope Stratigraphy Across the Continental Paleocene-Eocene Boundary on Polecat Bench in the Northern Bighorn Basin, *in* Gingerich, P.D. ed., *Paleocene-Eocene Stratigraphy and Biotic Change in the Bighorn and Clarks Fork Basins, Wyoming*, University of Michigan Papers on Paleontology, 33, pp 73-88.

Briggs, John C. *Biogeography and Plate Tectonics*. Developments in Palaeontology and Stratigraphy series, volume 10. Elsevier Science Publishers B.V. Amsterdam, The Netherlands, 1987.

Buchardt, Bjorn, 1978. Oxygen Isotope Palaeotemperatures from the Tertiary Period in the North Sea: *Nature*, Vol. 275, pp 121-123.

Campbell, David C., Campbell, Matthew R., 2003. Biotic Patterns in Eocene-Oligocene Molluscs of the Atlantic Coastal Plain, U.S.A. *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., *From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition*: New York, Columbia University Press, pp 341-353.

- Carter, Burchard D., 2003. Diversity Patterns in Eocene and Oligocene Echinoids of the Southeastern United States *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition: New York, Columbia University Press, pp 354-385.
- Cavelier, Claude, Chateauneuf, Jean-Jacques, Pomerol, Charles, Rabussier, Dominique, Renard, Maurice and Colette Vergnaud-Grazzini, 1981. The Geological Events at the Eocene/Oligocene Boundary: Palaeogeography, Palaeoclimatology, Palaeoecology, Vol. 36, pp 223-248.
- Cerling, Thure, E., 1984. The Stable Isotopic Composition of Modern Soil Carbonate and its Relationship to Climate: Earth and Planetary Science Letters, Vol 71, pp 229-240.
- Cerling, Thure E., 1991. Carbon Dioxide in the Atmosphere: Evidence From Cenozoic and Mesozoic Paleosols: American Journal of Science, Vol. 291, pp. 377-400.
- Cerling, T. E., Quade, J., Wang, Y., Bowman, J. R., 1989. Carbon Isotopes in Soils and Paleosols as Ecology and Palaeoecology Indicators: Nature, Vol 341, pp 138-139.
- Collinson, M.E., Fowler, K., Boulter, M.C., 1981. Floristic Changes Indicate a Cooling Climate in the Eocene of Southern England: Nature, Vol. 291, pp 315-317.
- DeConto, Robert M., Pollard, David, 2003. Rapid Cenozoic Glaciation of Antarctica Induced by Declining Atmospheric CO₂: Nature, Vol 421, pp 245-249.
- DeConto, Robert M., Pollard, David, Wilson, Paul A., Palike, Heiko, Lear, Caroline H., Pagani, Mark, 2008. Thresholds for Cenozoic Bipolar Glaciation: Nature, Vol. 455, pp 652-657.
- Denniston, Rhawn F., DuPree, Michelle, Dorale, Jeffrey A., Asmerom, Yemane, Polyak, Victor J., Carpenter, Scott J., 2007. Episodes of Late Holocene Aridity Recorded by Stalagmites from Devil's Icebox Cave, Central Missouri, USA: Quaternary Research, vol. 68, pp. 45-52.
- Diester-Haass, Liselotte, Zahn, Rainer, 1996. Eocene-Oligocene Transition in the Southern Ocean: History of Water Mass Circulation and Biological Productivity: Geology, Vol. 24, pp 163-166.
- Dockery, David T. III, Lozouet, Pierre, 2003. Molluscan Faunas Across the Eocene/Oligocene Boundary in the North American Gulf Coastal Plain, with Comparisons to Those of the Eocene and Oligocene of France *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition: New York, Columbia University Press, pp 303-340.
- Dwyer, Gary S., 2000. Unraveling the Signals of Global Climate Change: Science, Vol. 287, pp. 246-247.
- Echols, R.J., Armentrout, J.M., Root, S.A., Fearn, L.B., Cooke, J.C., Rodgers, B.K., and Thompson, P.R., 2003. Sequence Stratigraphy of the Eocene/Oligocene Boundary Interval: Southeastern Mississippi, *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition: New York, Columbia University Press, pp 189-222.

- Eldrett, J.S., Harding, I.C., Wilson, P.A., Butler, E., and Robers, A.P., 2007. Continental Ice in Greenland During the Eocene and Oligocene: *Nature*, Vol. 466, pp 176-179.
- Eshet, Yoram, Druckman, Yehezkeel, Cousminer, Harold L., Habib, Daniel, Drugg, Warren S., 1988. Reworked Palynomorphs and their use in the Determination of Sedimentary Cycles: *Geology*, Vol. 16, pp 662-665.
- Evanoff, E., Prothero, D.R., Lander, R.H., 1992. Eocene-Oligocene Climatic Changes in North America: The White River Formation near Douglas, East-Central Wyoming, *in* Prothero, Donald R., Berggren, William A., eds., *Eocene-Oligocene Climatic and Biotic Evolution*: New Jersey, Princeton University Press, pp 116-130.
- Evans, James E., 1999. Recognition and Implications of Eocene Tufas and Travertines in the Chadron Formation, White River Group, Badlands of South Dakota: *Sedimentology*, Vol 46, pp 771-789.
- Evans, James E., Welzenbach, Linda C., 2000. Lacustrine Limestones and Tufas in the Chadron Formation (Late Eocene), Badlands of South Dakota, U.S.A in E.H. Gierlowski-Kordesch and K.R. Kelts, eds., *Lake Basins Through Space and Time: AAPG Studies in Geology* 46, pp 349-358.
- Exon, Neville, Kennett, Jim, Molone, Mitch; Brinkhuis, Henk; Chaproniere, George; et al., 2002. Drilling Reveals Climatic Consequences of Tasmanian Gateway Opening: *EOS, Transactions, American Geophysical Union*, Vol. 83, No. 23, p 253-264
- Fawcett, Peter J., Boslough, M.B.E., 2002. Climatic Effects of an Impact-Induced Equatorial Debris Ring: *Journal of Geophysical Research*, Vol. 107, Issue 15.
- Finnigan GasBench II Operating Manual, Thermo Electron Corporation: Bremen, Germany, 2004.
- Fluegeman, Richard, 1996. Preliminary Paleontological Report on the Foraminifera of the Mossy Grove Core, Hinds County, Mississippi: *Mississippi Geology*, Vol. 17, pp 9-15.
- Fluegeman, Richard, 2003. Late Eocene-Early Oligocene Benthic Foraminifera in the Gulf Coastal Plain: Regional vs. Global Influences, *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., *From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition*: New York, Columbia University Press, pp 283-293.
- Fluegeman, Richard, 2009. Late Eocene and Early Oligocene Benthic Foraminiferal Paleocology and Sequence Stratigraphy in the Eastern Gulf Coastal Plain, U.S.A., *in* Demchuk, M.T. and Gary, A., eds., *Geologic Problem Solving with Microfossils: Society of Economic Paleontologists and Mineralogist (SEPM) Special Publication no. 93*, pp 293-307.
- Fluegeman, Richard H., Grigsby, Jeffrey D., Hurley, John V., Eocene-Oligocene Greenhouse to Icehouse Transition on a Subtropical Clastic Shelf: The Jackson-Vicksburg Groups of the Eastern Gulf Coastal Plain of the United States, *in* Koeberl, C., and Montanari, A., eds. *The Late Eocene Earth- Hothouse, Icehouse and Impacts: Geological Society of America Special Paper 452*, pp. 261-277.
- Hartenberger, Jean-Louis, 1998. An Asian Grande Coupure: *Nature*, Vol. 394, pp 321.

- Hurley, John V., Fluegeman, Richard H., 2003. Late Middle Eocene Glacioeustasy: Stable Isotopes and Foraminifera from the Gulf Coastal Plain, *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., *From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition*: New York, Columbia University Press, pp 223-231.
- Hutchison, J.H., 1982. Turtle, Crocodylian, and Champsosaur Diversity Changes in the Cenozoic of the North-Central Region of Western United States: *Palaeogeography, Palaeoclimatology, Palaeoecology*, Vol. 37, pp 149-164.
- Ivany, Linda C., Patterson, William P., Lohmann, Kyger C., 2000. Cooler Winters as a Possible Cause of Mass Extinctions at the Eocene/Oligocene Boundary: *Nature*, Vol. 407, pp 887-890.
- Ivany, Linda C., Nesbitt, Elizabeth A., Prothero, Donald R., 2003. The Marine Eocene-Oligocene Transition: A Synthesis *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., *From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition*: New York, Columbia University Press, pp 522-534.
- Jacobs, Bonnie F., 2004. Palaeobotanical Studies from Tropical Africa: Relevance to the Evolution of Forest, Woodland and Savannah Biomes: *Philosophical Transactions of the Royal Society of London B*, Vol. 359, pp 1573-1583.
- Kennett, J.P., von der Borch, C., Baker, P.A., Barton, C.E., Boersma, A., Cauler, J.P., Dudley, W.C. Jr., Gardner, J.V., Jenkins, D.G., Lohman, W.H., Martini, E., Merrill, R.B., Morin, R., Nelson, C.S., Robert, C., Srinivasan, M.S., Stein, R., Takeuchi, A., & Murphy, M.G., 1985. Palaeotectonic Implications of Increased Late Eocene-Early Oligocene Volcanism from South Pacific DSDP Sites: *Nature*, Vol 316 pp 507-511.
- Kraus, Mary J., 1997. Lower Eocene Alluvial Paleosols: Pedogenic Development, Stratigraphic Relationships, and Paleosol/Landscape Associations: *Palaeogeography, Palaeoclimatology, Palaeoecology*, Vol. 129, pp 387-406.
- Lear, C.H., Elderfield, H., Wilson, P.A., 2000. Cenozoic Deep-Sea Temperatures and Global Ice Volumes from Mg/Ca in Benthic Foraminiferal Calcite: *Science*, Vol. 287, pp 269-272.
- McBride, Earle F., Lindemann, William L., Freeman, Paul S., 1968. Lithology and Petrology of the Gueydan (Catahoula) formation in South Texas. Report of Investigations No. 63; Bureau of Economic Geology, the University of Texas at Austin, Austin, TX.
- Meng, Jin, McKenna, Malcolm C., 1998. Faunal Turnovers of Palaeogene Mammals from the Mongolian Plateau: *Nature*, Vol. 394, pp 364-367.
- Miller, Kenneth G., 1992. Middle Eocene to Oligocene Stable Isotopes, Climate, and Deep-Water History: The Terminal Eocene Event?, *in* Prothero, Donald R., Berggren, William A., eds., *Eocene-Oligocene Climatic and Biotic Evolution*: New Jersey, Princeton University Press, pp 131-159.
- Miller, Kenneth G., Browning, James V., Aubry, Marie-Pierre, Wade, Bridget S., Katz, Miriam E., Kulpecz, Andrew Al., Wright, James D., 2008. Eocene-Oligocene Global Climate and Sea-Level Changes: St. Stephens Quarry, Alabama: *Geological Society of America Bulletin*, Vol., 120, pp 34-53.

Nikiforoff, C.C., 1949. Weathering and Soil Evolution: *Soil Science*, Vol. 67, issue 2, pp 219.

National Park Service, Badlands National Park:

www.nps.gov/badl/planyourvisit/weather.htm. Site last modified May 1, 2009.

O'Keefe, John A., 1980. The Terminal Eocene Event: Formation of a Ring System Around the Earth?: *Nature*, Vol. 285, pp. 309-311.

Owen, Robert M., Rea, David K., 1985. Sea-Floor Hydrothermal Activity Links Climate to Tectonics: The Eocene Carbon Dioxide Greenhouse: *Science*, Vol. 227, pp 166-169.

Poag, C. Wylie, Mankinen, Edward, Norris, Richard D., 2003. Late Eocene Impacts: Geologic Record, Correlation, and Paleoenvironmental Consequences *in* Prothero, D., Ivany, L., and Nesbitt, E., eds., *From Greenhouse to Icehouse The Marine Eocene-Oligocene Transition*: New York, Columbia University Press, pp 495-509.

Prothero, Donald R., 1994. *The Eocene-Oligocene Transition: Paradise Lost*: New York, Columbia University Press.

Prothero, D.R., 1999. Does Climatic Change Drive Mammalian Evolution?: *GSA Today*, Vol. 9., No. 9, pp 1-7.

Prothero, Donald R., Swisher, Carl C. III, 1992. Magnetostratigraphy and Geochronology of the Terrestrial Eocene-Oligocene Transition in North America *in* Prothero, Donald R., Berggren, William A., eds., *Eocene-Oligocene Climatic and Biotic Evolution*: New Jersey, Princeton University Press, pp 46-73.

Prothero, D. R., Whittlesey, K.E., 1998. Magnetic Stratigraphy and Biostratigraphy of the Orellan and Whitneyan Land Mammal "Ages" in the White River Group, *in* Terry, D., LaGarry, H., Hunt, R., eds., *Depositional Environments, Lithostratigraphy, and Biostratigraphy of the White River and Arikaree Groups (Late Eocene to Early Miocene, North America)*: Geological Society of America Special Paper 325, pp 93-116.

Retallack, Greg J., 1983. A Paleopedological Approach to the Interpretation of Terrestrial Sedimentary Rocks: The Mid-Tertiary Fossil Soils of Badlands National Park, South Dakota: *Geological Society of America Bulletin*, Vol. 94, pp 823-840.

Retallack, Greg J., 1985. An Excursion Guide to Fossil Soils of the Mid-Tertiary Sequence in Badlands National Park, South Dakota *in* Marin, James E., ed., *DAKOTERRA*, Volume 2, Part 2: Fossiliferous Cenozoic Deposits of Western South Dakota and Northwest Nebraska, pp 277-301.

Retallack, Gregory, J., 1992. Paleosols and Changes in Climate and Vegetation Across the Eocene/Oligocene Boundary *in* Prothero, Donald R., Berggren, William A., eds., *Eocene-Oligocene Climatic and Biotic Evolution*: New Jersey, Princeton University Press, pp 382-398.

Retallack, G.J., 1998. Fossil Soils and the Completeness of the Rock and Fossil Records *in* Donovan, S.K. and Paul, C.R.C., eds., *The Adequacy of the Fossil Record*: Chichester, UK, John Wiley and Sons, pp 131-162.

Retallack, Gregory J., 2001. *Soils of the Past an Introduction to Paleopedology*, 2nd Edition: Oxford, Blackwell Science Ltd, 404 pages.

- Savin, Samuel M., 1980. Pre-Pleistocene Climates: *Nature*, Vol. 286, pp 553-554.
- Sloan, Lisa Cirbus, Barron, Eric J., 1992. Paleogene Climatic Evolution: A Climate Model Investigation of the Influence of Continental Elevation and Sea-Surface Temperature Upon Continental Climate *in* Prothero, Donald R., Berggren, William A., eds., *Eocene-Oligocene Climatic and Biotic Evolution*: New Jersey, Princeton University Press, pp 202-217.
- Stevenson, F. J., February 1969. Pedohumus: Accumulation and Diagenesis During the Quaternary: *Soil Science*, Vol. 107, No. 6, p 470-479
- Stoffer, Philip W., 2003. *Geology of Badlands National Park: A Preliminary Report*: USGS Open-File Report 03-35, Menlo Park, California, 60 pages.
- Stöffler, Dieter, Claeys, Phillippe 1987. Earth Rocked by Combination Punch: *Nature*, Vol. 388, pp 331-332.
- Swisher, Carl C. III, Prothero, D.R., 1990. Single-Crystal $^{40}\text{Ar}/^{39}\text{Ar}$ Dating of the Eocene-Oligocene Transition in North America: *Science*, Vol. 249, pp. 760-762.
- Tripati, Aradhna, Backman, Jan, Elderfield, Henry, Ferretti, Patrizia, 2005. Eocene Bipolar Glaciation Associated With Global Carbon Cycle Changes: *Nature*, Vol 436, pp 341-346.
- Tripati, Aradhna, Eagle, R., Morton, A., Dowdeswell, J., Atkinson, K., Bahe, Y., Dawber, F., Khadun, E., Shaw, R., Shorttle, O., and Thanabalasundaram, L., 2008. Evidence for Northern Hemisphere Glaciation Back to 44 Ma from Ice-Rafted Debris in the Greenland Sea: *Earth and Planetary Science Letters*, Vol. 265, pp 112-122.
- Vonhof, Hubert B., Smit, Jan, Brinkhuis, Henk, Montanari, Alessandro, Nederbragt, Alexander J., 2000. Global Cooling Accelerated by Early Late Eocene Impacts?: *Geology*, Vol. 28, no. 8, pp 687-690.
- Walker, Theodore R., March 1967. Formation of Red Beds in Modern and Ancient Deserts, *Geological Society of America Bulletin*, V. 78, p. 353-368.
- Zachos, James C., Breza, James R., Wise, Sherwood W., 1992. Early Oligocene Ice-Sheet Expansion on Antarctica: Stable Isotope and Sedimentological Evidence from Kerguelen Plateau, Southern Indian Ocean: *Geology*, Vol 20, pp 569-573.
- Zachos, James C., Stott, Lowell D., Lohmann, Kyger C., 1994. Evolution of Early Cenozoic Marine Temperatures: *Paleoceanography*, Vol. 9, pp.353-387.
- Zachos, James C., Opdyke, Bradley N., Quinn, Terrence M., Jones, Charles E., Halliday, Alex N., 1999. Early Cenozoic Glaciation, Antarctic Weathering, and Seawater $^{87}\text{Sr}/^{86}\text{Sr}$: is There a Link?: *Chemical Geology*, Vol. 161, pp 165-180.
- Zachos, James, Pagani, Mark, Sloan, Lisa, Thomas, Ellen, Billups, Katharina, 2001. Trends, Rhythms, and Aberrations in Global Climate 65 Ma to Present: *Science*, Vol 292, pp 686-693.
- Zanazzi, Alessandro, Kohn, Matthew J., MacFadden, Bruce J., Terry, Dennis O. Jr., 2007. Large Temperature Drop Across the Eocene-Oligocene Transition in Central North America: *Nature*, Vol. 445, pp 639-642.