Evidence for an Albian Hudson arm connection between the Cretaceous Western Interior Seaway of North America and the Labrador Sea

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ABSTRACT

Numerous researchers have alluded to the existence of a Cretaceous Hudson Arm connection between the Labrador Sea and the Western Interior Seaway of North America. However, the evidence for this marine connection has been circumstantial. In this paper we present sedimentary geochemical data that indicate a marine influence in the Albian Mattagami Formation of the Moose River basin, James Bay Lowlands, Ontario. The facies associations between dinoflagellate-bearing laminated mudstones, fluvial sandstones, and early pyrite mineralization are interpreted to indicate deposition in the central basin of an estuary. We use the facies association between the estuarine fill and coeval kaolinitic paleosols in the Moose River basin, and in similar deposits in Quebec and Labrador, to reconstruct a southern shoreline of the Albian Hudson Arm to the Cretaceous Western Interior Seaway. We suggest that development of the Hudson Arm connection between the Labrador Sea and the Cretaceous Western Interior Seaway may be related to a regional extensional regime associated with rifting between Labrador and Greenland, and the passage of eastern North America over Cretaceous hotspots.

Keywords: Cretaceous, geochemistry, Hudson Bay, Mattagami Formation, Western Interior Seaway.

INTRODUCTION

The paleogeography of the Cretaceous Western Interior Seaway is well established (e.g., McGookey et al., 1972; Reynolds and Dolly, 1983; Merewether and Cobban, 1986; Roberts and Kirschbaum, 1995; and many others), yet uncertainty remains with regard to the so-called Hudson Arm of the seaway. Jeletzky (1971) first postulated the existence of a Turonian highstand connection between the Cretaceous Western Interior Seaway and Baffin Bay based on a study of ammonites (Fig. 1). Williams and Stelck (1975, p. 10) proposed a connection for much of the Late Cretaceous and stated that ''the interior seaway extended into the Hudson Bay area in Turonian time.''

Ziegler and Hulver (1989) inferred a Cenomanian connection between the Cretaceous Western Interior Seaway and Labrador Sea through the Williston and Hudson Bay basins in their free air gravity reconstruction of the Canadian shield. They noted glacial tills bearing a marine fauna of Turonian Greenhorn Formation affinity found in Manitoba, and concluded that ice flow directions and the distance from the Greenhorn outcrop area to the south were suggestive of a source in the Hudson Bay region (Ziegler and Hulver, 1989). On the basis of seismic data, Sanford and Grant (1990) interpreted 150 m of strata on the seafloor of Hudson Bay as Cretaceous sandstone and shale, and mapped a ring-shaped distribution encompassing much of the central part of the basin (Fig. 1). The age determination was based on a mixed assemblage of palynomorphs from a grab sample of the seafloor (McGregor, 1987), later defined as Aptian to Cenomanian (Sanford and Grant, 1990).

The presence of the *Muderongia asymmetrica* dinoflagellate assemblage in Albian rocks from the Labrador Shelf, Canadian Arctic, and the Cretaceous Western Interior Seaway was used to suggest a connection between these marine seaways (Williams et al., 1990). Furthermore, Albian marine dinoflagellates found in the Northwest Territories, Canada, suggest that the Cretaceous Western Interior Seaway was more extensive than has previously been thought (Nassichuk and McIntyre, 1995). Diner et al. (1996) recognized Cenomanian–Turonian benthic foraminiferal taxa from the Manitoba Escarpment that are uncharacteristic of the faunas of the North American Western Interior, and included species from coeval strata in Europe. They presented this information as evidence for a connection between the Cretaceous Western Interior Seaway and European epeiric seas (Diner et al., 1996).

Although the paleogeography of the various hypothesized connections is not clear due to a lack of outcrops and the absence of marine invertebrate faunas such as inoceramids and ammonites, a connection across Hudson Bay provides the simplest reconstruction. Marine conditions in the Labrador Sea were established during the Albian–Turonian sea-level highstand, extending into the southeast Baffin shelf (Gradstein et al., 1990). Embry (1991) interpreted late Aptian to early Cenomanian rifting in the Labrador Sea–Baffin Bay region as the link between the Sverdrup basin (Canadian Arctic Archipelago) and the Atlantic Ocean, but stated that an uplift separated the Sverdrup basin from the proto–Arctic Ocean. We suggest that the combination of faunal ties between the Cretaceous Western Interior Seaway and the proto–North Atlantic and the lack of an Albian–Turonian physical connection between the Labrador Sea and the Arctic Ocean provide an intriguing argument for implicating a connection between the Atlantic Ocean and the Cretaceous Western Interior

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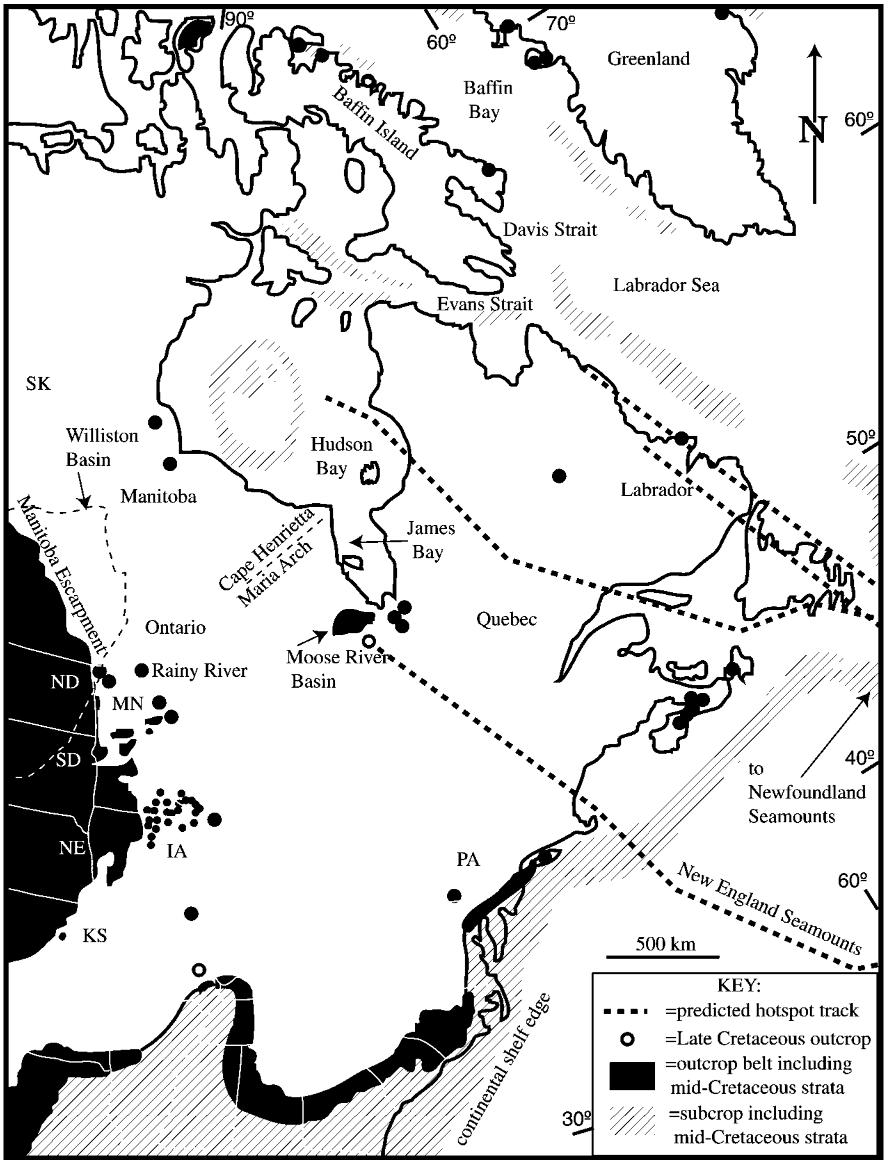

Figure 1. Geologic map of Cretaceous strata in most of eastern North America (from Balkwill et al., 1983; Blais, 1959; Burden and Langille, 1991; Burden and Holloway, 1985; Cook and Bally, 1972; Escher and Pulvertaft, 1995; Fuller, 1961; Gohn, 1988; Grant and McAlpine, 1990; Henderson et al., 1976; King and McMillan, 1975; Ludvigson and Witzke, 1996; McFarlan and Menes, 1991; Miall et al., 1980; Norris, 1993; Norris and Zippi, 1991; Remick et al., 1963; Sanford and Grant, 1990; Sohl et al., 1991; Tschudy, 1965; Umpleby, 1979; Wade and MacLean, 1990; Williams, 1948; Ziegler and Rowley, 1998; and Zippi and Bajc, 1990), including hypothesized hotspot tracks associated with the New England and Newfoundland Seamounts (Duncan, 1984). Note that black dots designate small outliers from the mid-Cretaceous outcrop belt.

Seaway through the Hudson Bay region. Thus far we have presented previously compiled circumstantial evidence to infer the existence of this connection during the Cenomanian–Turonian global sea-level highstand. However, the apparent paucity of in situ evidence for this Hudson Arm connection has fostered skepticism. In this paper we describe in situ evidence for marine influence in late Albian strata of the Moose River basin, James Bay Lowlands, Ontario, Canada (Fig. 1), previously interpreted as exclusively nonmarine in origin. We suggest that marine influence in these strata is most easily explained by a marine connection through the present-day Hudson Bay region.

Moose River Basin

The Moose River basin is located in the James Bay Lowlands of Ontario, Canada, and is one of two sedimentary basins on the Hudson platform. The Moose River basin contains <1000 m of Paleozoic–Mesozoic strata, and is separated from the Hudson Bay basin to the north by the Cape Henrietta Maria arch (Fig. 1; Telford, 1991). The basin formed by gravity sag and block faulting of the stable Precambrian shield (Sanford et al., 1985).

Mattagami Formation

Cretaceous strata in the Moose River basin compose the Mattagami Formation (Keele, 1920; Dyer, 1928). The formation consists of varicolored mudrock, with quartz sandstone, conglomerate, and lignite (Try et al., 1984). Try et al. (1984) interpreted the Mattagami Formation as deposited in a ''high-constructive segment of a major large scale river system which drained an extensive tract of the Canadian Shield.'' (Try et al., 1984, p. 345). They named this river the Esoom river, and showed it flowing in a northwesterly direction and discharging into the Cretaceous Western Interior Seaway (Try et al., 1984). Palynological studies of the Mattagami Formation (e.g., Hopkins and Sweet, 1976; Norris et al., 1976; and others) culminated in the zonation of Norris and Zippi (1991). In their study, the Mattagami Formation was interpreted to range from the Aptian–early Albian to the late Albian, with age determinations easier toward the top of the section because more comparative Albian–Cenomanian studies are available than for earlier Cretaceous stages (Norris and Zippi, 1991). Telford et al. (1975, p. 18) stated that Albian acritarchs in the Mattagami Formation ''might suggest marine influence.'' Laminated mudstones interpreted as lacustrine (Try et al., 1984) were reported to contain dinoflagellates (Norris and Zippi, 1986). Long (1991, p. 82) considered that the dinoflagellates may have been deposited by winds ''or on the feet of migratory birds.'' However, Norris and Zippi (1991, p. 122) stated that although a freshwater origin for the Mattagami Formation is supported by the lack of marine fossils and the low-diversity dinocyst assemblages, dinoflagellates imply brackish-marine water, and ''may indicate a connection to a marine body of water.'' Zippi (1998) suggested that cysts in the laminated mudstones are one of the earliest freshwater dinoflagellates. The results of our geochemical analyses of these dinoflagellate-bearing laminated mudstones from the Mattagami Formation are the focus of this paper.

DINOFLAGELLATE-BEARING MUDSTONES OF THE MATTAGAMI FORMATION

Geological exploration of the James Bay lowlands has been ongoing for more than 100 yr. In the process, a number of boreholes have been drilled. Among these are the Ontario Geological Survey (OGS) 84-06 and 84-08 cores, housed at the OGS Core Warehouse in Timmins, Ontario. The cores were drilled ~6 km apart in the Cretaceous subcrop of the Moose River basin. Both cores penetrated the Mattagami Formation and were included in the palynological characterization of Norris and Zippi (1991) and Zippi (1998) (discussed later). Both cores contain intervals of ~10 m of red-mottled, rooted, sphaerosiderite-bearing, kaolinitic mudrock paleosols interbedded with dark gray, laminated, organic-rich mudstones.

It is evident from our lithologic descriptions and those of the OGS (Long, 1991) and Zippi (1998) that the paleosol and laminated mud-rock horizons are correlatable between the cores (Fig. 2). The study interval in the OGS 84-08 core consists of a basal ~1 m, red-mottled, rooted, sphaerosiderite-bearing, sandy, kaolinitic mudrock paleosol overlain by two lignites and sandstone. This interval is overlain by ~2 m of dark gray, dinoflagellate-bearing laminated mudstone, overlain by a 1 m sandy kaolinitic mudrock paleosol. An ~2.5 m black dinoflagellate-bearing laminated mudstone with interbedded sand is above this paleosol, and is overlain by a >2 m kaolinitic mudrock paleosol. In the OGS 84-06 core, an interval of black dinoflagellate-bearing laminated mudstone with sandstone interbeds is present from 123 to ~119 m in the core. This unit in the OGS 84-06 core is correlated with the lowermost dinoflagellate-bearing laminated mudstone in the OGS 84-08 core, and is overlain by a paleosol that is correlated with the middle paleosol in the 84-08 core. An ~4 m dinoflagellate-bearing laminated mudstone, correlated with the upper black mudstone in the OGS 84-08 core, overlies the paleosol at ~117 m in the OGS 84-06 core. The upper meter of the study interval in the OGS 84-06 core contains mostly sandstone.

RESEARCH METHODS

We visited the Timmins core facility in 1997 to sample the dinoflagellate-bearing laminated mudstone and kaolinitic paleosol horizons at 0.33 m intervals through both OGS cores. The samples were evaluated at the Sedimentary Geochemistry Lab at Pennsylvania State University, where they were crushed and sieved through a −100 mesh screen. Each sample was subjected to carbon coulometric titration (Engleman et al., 1985) for percent inorganic and percent total carbon (total organic carbon, %TOC, was calculated by difference), $\%FeS_2$ analysis by coulometric titration of H_2S using a sulfur coulometer (Wilken and Barnes, 1997), and Rock-Eval pyrolysis (Peters, 1986) for a measure of hydrogen and oxygen richness. The Rock-Eval method uses programmed heating of a sample to measure concentrations of free and adsorbed hydrocarbons (S_1), pyrolytic hydrocarbons from kerogen degradation (S_2), and CO_2 generated from kerogen degradation (S_3 peak). The S_2 and S_3 peak areas normalized to %TOC provide a hydrogen index (HI) and oxygen index (OI) that correlate to atomic H:C and O:C ratios used to differentiate between organic matter types (Robert, 1985; Peters, 1986; discussed in the following). The results of these analyses are shown in Figures 3 and 4. Minor discrepancies between geochemical sample depths and the lithologic descriptions (Fig. 2) are attributed to incomplete core recovery.

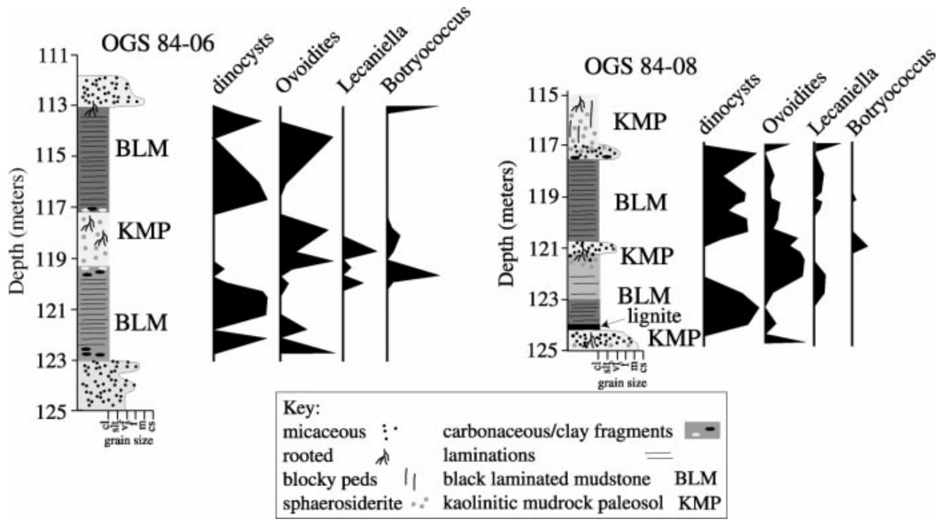

Figure 2. Lithologic core descriptions and select palynomorph distributions for the study interval in the Albian Mattagami Formation, Moose River basin, Ontario, Canada (modified from Zippi, 1998). BLM—black laminated mudstones; KMP—kaolinitic mudrock paleosol.

RESULTS

The paleosol units contain the highest carbonate content ($\%CaCO_3$) in the study interval in both cores, ranging from near 0% to ~10%, and the basal paleosol in both cores has the highest $\%CaCO_3$ value; OI values from Rock-Eval pyrolysis display a similar distribution (Figs. 3 and 4). The %TOC values mostly range from 0% to ~8%, though two lignites at the base of OGS 84-08 contain ~40% and ~50%TOC (Figs. 3 and 4). Neglecting the lignites, the highest %TOC values are in the dinoflagellate-bearing laminated mudstones. HI values covary with %TOC, so the highest HI values are found in the dinoflagellate-bearing laminated mudstones (Figs. 3 and 4). FeS_2 values are mostly <0.5%, although elevated values (to ~2.2% in OGS 84-06) are associated with the dinoflagellate-bearing laminated mudstones and immediately beneath the lower dinoflagellate-bearing laminated mudstone in OGS 84-08 (Figs. 3 and 4).

DISCUSSION

We explain higher values for $\%CaCO_3$ in paleosols in the Mattagami Formation with the observation that some sphaerosiderites in the paleosols are enveloped by blocky calcite spar that overprinted the sphaerosiderites during a later episode of carbonate diagenesis. The relatively higher OI values in the paleosols are likely due to the presence of oxidized organic matter in the paleosols, although the thermal decomposition of iron carbonate minerals cannot be ruled out (Es-

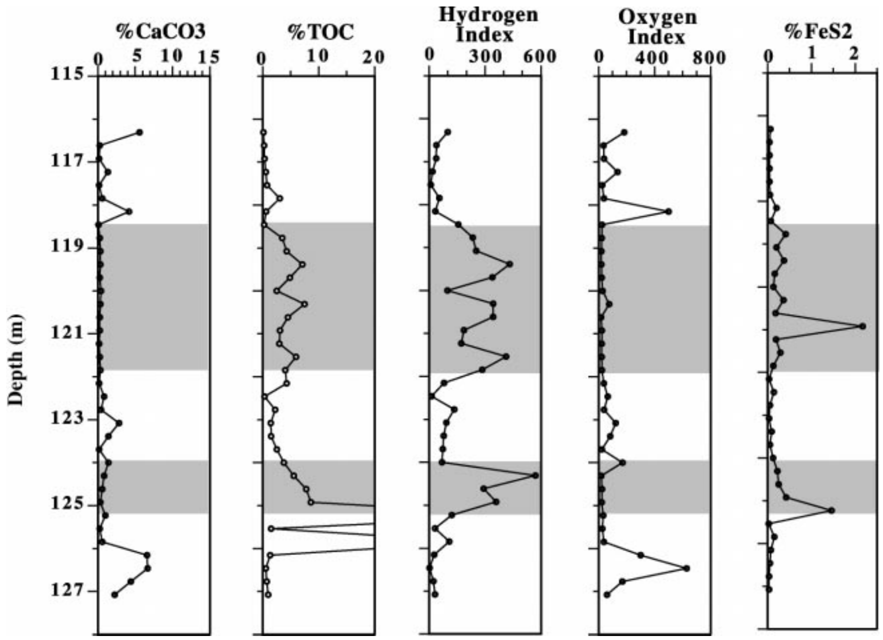

Figure 3. Distribution of $\%CaCO_3$, %TOC (total organic carbon), hydrogen and oxygen index, and $\%FeS_2$ for the Ontario Geological Survey 84–08 core, Albian Mattagami Formation, Moose River basin, Ontario, Canada; shaded areas mark dinoflagellate-bearing laminated mudstones.

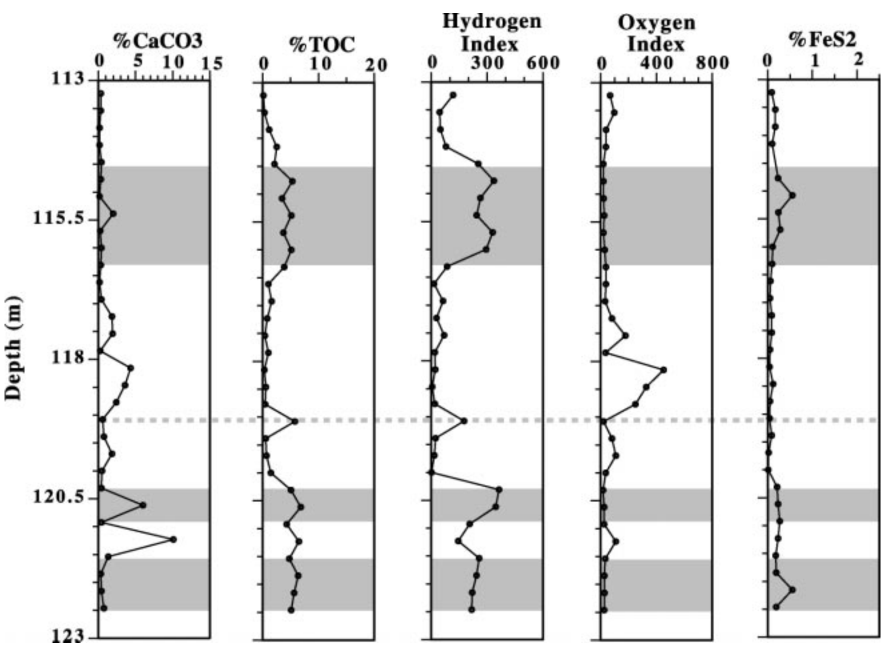

Figure 4. Distribution of $\%CaCO_3$, %TOC (total organic carbon), hydrogen and oxygen index, and $\%FeS_2$ for the Ontario Geological Survey 84–06 core, Albian Mattagami Formation, Moose River basin, Ontario, Canada; shaded areas and dashed line mark dinoflagellate-bearing laminated mudstones.

pitalie et al., 1977; Katz, 1983). The focus of the remainder of the paper is on the relationship between organic matter quantity and type and pyrite in the Mattagami Formation as evidence for a Hudson Arm marine connection between the proto–Atlantic Ocean and the Cretaceous Western Interior Seaway.

Organic Matter

The modified van Krevelen diagram in Figure 5A shows that the distribution of organic matter types in the study interval may include types I, II, III, and IV. Type I organic matter is generally considered to be derived from a lacustrine algal source, whereas type II is often considered to have marine algal affinities. Type III organic matter is derived from higher plants with terrestrial affinities, or can be oxidized type I or II organic matter, and type IV represents highly oxidized organic matter (Robert, 1985; Peters, 1986). With this in mind, a review of Figures 3 and 4 shows that the lignites near the base of OGS 84-08 are composed of type III organic matter, although most of the type III and IV organic matter in the cores is found in paleosols and sandstone. Type I and II organic matter is found in the laminated mudstones. Unfortunately, the plot does not provide direct insight to a marine or lacustrine origin for the organic matter.

Langford and Blanc-Valleron (1990) suggested an alternate approach to assessing organic matter type using Rock-Eval pyrolysis data. Their method uses raw S_2values (mg hydrocarbons/g rock) from Rock-Eval pyrolysis without normalizing to %TOC. Their work shows that the pyrolytic hydrocarbon yields of type I, II, and III organic matter can be separated on an S_2 vs. %TOC plot (Langford and Blanc-Valleron, 1990). Figure 5B is a plot of S_2 vs. %TOC for the Mattagami Formation samples, showing that the organic matter is best characterized as type II (marine algal) and type III (terrestrial). This suggests that the dinoflagellate-bearing laminated mudstones in the Mattagami Formation may contain marine algal organic matter.

The geochemically based identification of marine organic matter associated with the dinoflagellate-bearing laminated mudstones is inconclusive, and contradicts the conclusions of earlier workers. For example, Zippi (1998, p. 1) clearly stated that ‘‘a diverse collection of fossil algae with botanical affinities to modern freshwater algae, were recovered from the Mattagami Formation.’’ However, other researchers (e.g., Bint, 1986; MacRae, 1992; Nunez-Betelu, 1994) noted the dominant Mattagami Formation dinoflagellate genus *Nykterïcysta* in coeval marine-influenced successions. In the following sections, we explore the record of pyrite as an indicator of marine influence in the Mattagami Formation.

Organic Matter and Pyrite

A review of Figures 3 and 4 shows that the highest values of %TOC, HI, and $\%FeS_2$ are found in black laminated mudstones in the Mattagami Formation. The covarying relationship between high HI values and high total organic carbon (disregarding the lignites at the base of OGS 84-08) in the formation is shown in Figure 5, C and D. Given the palynological information of Zippi (1998), these figures show that dinoflagellate productivity led to the deposition and preservation of labile organic matter in the black laminated mudstones. In Figure 5, E and F, higher values of $\%FeS_2$ are associated with higher values of total organic carbon. The organic carbon to pyritic sulfur (C/S) line with a slope of 5 is used to delineate marine versus freshwater environments of deposition (Berner and Raiswell, 1984); a C/S = 2 line for Upper Cretaceous marine shales is also included on the plots (Raiswell and Berner, 1986). The majority of the data are within the interpreted freshwater region on the plots, although the highest $\%FeS_2$ value observed in the study (2.17), and many of the lower values, are within the interpreted marine region. A linear trend with a zero intercept is observable in the data set, which may be suggestive of marine conditions with pyrite formation limited by organic carbon availability (Berner and Raiswell, 1983; Leventhal, 1983). Scatter in the plot may be attributed to euxinic environments in which pyrite formation occurred in the water column, with no systematic relationship between pyrite and organic carbon (Leventhal, 1983), or to iron limitation in a high-productivity zone (Beier and Hayes, 1989).

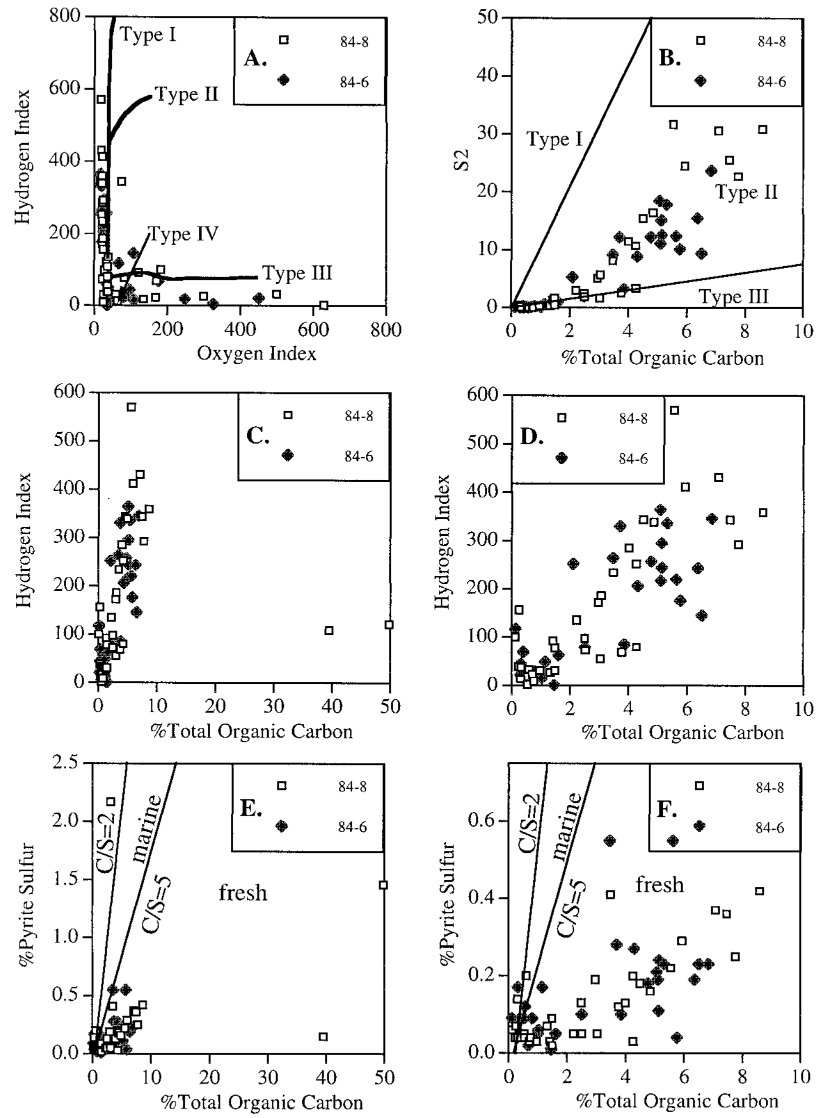

Figure 5. Plots of geochemical parameters displaying the relationships between %TOC (total organic carbon), organic matter type, and $\%FeS_2$ for the Mattagami Formation, Moose River basin, Ontario, Canada. 84-8 and 84-6 in the legends refer to Ontario Geological Survey (OGS) cores 84-8 and 84-6. (A) Organic matter types plotted on a modified Van Krevelen diagram. (B) Organic matter types as determined using the method of Langford and Blanc-Valleron (1990). (C) Hydrogen index versus %TOC. (D) Same as C, with focus on TOC <10%. (E) %pyrite versus %TOC with C/S ratios delineating fresh versus marine environments of formation. (F) Same as E, with focus on TOC <10%.

Controls on Pyrite Formation

In general, large quantities of iron sulfide can form in sulfate-rich marine systems, whereas iron sulfide formation is often limited by low sulfate availability in freshwater settings (Davison, 1988). Iron sulfides precipitate only under very reducing conditions, or from solutions with unusually high sulfide concentrations (Krauskopf, 1979). In freshwater sediments, pyrite appears to form in low concentrations at redox boundaries, although most of the sulfur in these systems is bound with organic matter (Davison, 1988). However, because the formation of pyrite in sediments is limited by the availability of labile organic matter, the sulfate concentration in the overlying water, and the availability of reduced iron (Berner and Raiswell, 1984), C/S ratios may be too simplistic for understanding paleosalinities. High %TOC and HI values in the Mattagami Formation demonstrate an ample supply of labile organic matter as an energy source for bacterial sulfate reducers, and Hamblin (1982) interpreted hornblende as an iron source for pyrite formation in the formation. The abundance of hematite and siderite observed in hand samples and thin sections also suggests an abundant iron supply. Because pyrite contents are generally low given the apparent ample iron and organic carbon supply, a restricted supply of sulfate for pyrite formation provides a reasonable explanation for low $\%FeS_2$ values in the Mattagami Formation. However, in a petrographic study of Mattagami Formation lignites, pyrite contents averaging 2% and as high as 3.5% were observed (Brown et al., 1986). In a study of Holocene marine and nonmarine peats in the southern United States, pyrite in marine peats approached 10%, while in freshwater peats sulfur values were <0.2% (Casagrande et al., 1977). Williams and Keith (1963) showed pyrite content in the Lower Kittanning coal of Pennsylvania as related to overlying marine shales. The $\%FeS_2$ values for Mattagami Formation lignites are suggestive of a marine-influenced setting.

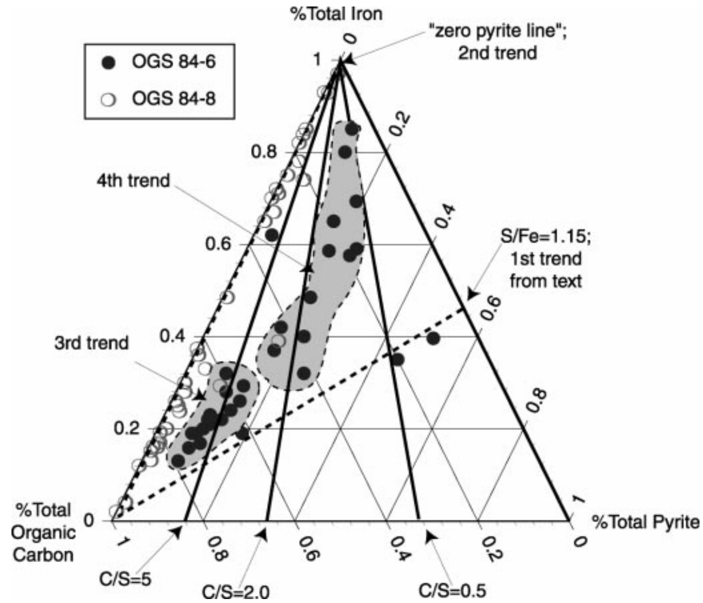

Figure 6. Fe-S-OC ternary diagram for samples from the Albian Mattagami Formation, Moose River basin, Ontario, Canada. Four trends are visible: (1) several points lie on the stoichiometric pyrite line at S/Fe = 1.15; (2) most of the data plot along or near the zero pyrite sulfur line, likely representing a freshwater environment of formation; (3) a data cluster on or to the left of the C/S = 5 line represents iron and sulfur-limited systems; and (4) the remainder of the data are between the C/S = 5 and C/S = 0.5 lines, and above the S/Fe = 1.15 line, representing pyrite formation in an organic-carbon–limited system. C/S = 2 line for Upper Cretaceous shale is from Raiswell and Berner (1986), and C/S = 5 line delineating fresh versus marine systems is from Berner and Raiswell (1984).

Dean and Arthur (1989) used ternary diagrams for understanding the Fe-S-OC system; the relationship between the variables is independent of dilution since ternary diagrams emphasize relative concentrations and ratios. The Fe-S-OC ternary diagram presented in Figure 6 provides further insights into the controls on pyrite formation in the Mattagami Formation. Four trends, or data clusters on the diagram, are worthy of note. First, several data points lie on the stoichiometric pyrite line labeled S/Fe = 1.15, emanating from the organic carbon apex of the diagram. The two points on the pyrite line that are closest to the pyritic sulfur axis are from the uppermost paleosol horizon in the OGS 84-06 core. The two points on the line closest to the TOC apex are from black laminated mudstones, and contain some of the highest %FeS_2 contents (~0.6% and ~1.5%) in the samples. Second, most of the data obtained from the OGS 84-08 core plots along or near the zero pyritic sulfur line. All of the points described on this line are greater than C/S = 5. Therefore, using the Berner and Raiswell (1984) scheme, those samples may represent freshwater environments of deposition. Most of the data closest to the total Fe apex were obtained from black laminated mudstones deposited in an environment with an abundant substrate of labile organic matter (high HI and %TOC) and iron, indicating that pyrite formation in this setting was mostly sulfur limited. Samples from the paleosols and sandstones in the OGS 84-08 core are closest to the TOC apex. The upper and lower paleosols contain small quantities of refractory organic matter, suggesting that pyrite formation in these settings was limited by a paucity of sulfur and labile organic matter. Third, a cluster of laminated mudstone data from the OGS 84-06 core plots between the constant S/Fe line and the zero sulfur line in the region of the diagram closest to the TOC apex. These data trend subparallel to the constant S/Fe line, and include one of the highest %FeS_2 contents (~0.6%). Note that many of these cluster data plot close to the constant C/S = 5 line. Pyrite formation in these samples was mostly iron limited, although at times sufficient iron may have been present, but diluted by an abundant supply of labile organic matter. In either case, the limitations on pyrite formation kept these samples from attaining an S/Fe = 1.15, although transformation to pyrite approached completion.

The fourth trend observable on the Fe-OC-S diagram involves the samples that are above the stoichiometric pyrite line, and between lines of constant C/S = 5 and C/S = 0.5 (Fig. 6). These samples mostly represent paleosol horizons from the OGS 84-06 core, although they include a laminated mudrock with the highest %FeS_2 content (~2.2%) in any of the Mattagami Formation samples. Because all of these paleosol samples contain low %TOC with low HI values, we suggest that pyrite formation in these samples was organic-carbon limited. The laminated mudrock with the highest %FeS_2 content contains some of the lowest %TOC and HI values for the laminated mudrock facies, indicating that pyrite formation in this setting may also have been carbon limited.

Timing of Pyrite Formation

Brown et al. (1986) determined that iron sulfide is the dominant mineral in Mattagami Formation lignites, and that pyrite in wood pores could have formed syngenetically or epigenetically, but pyrite casts of wood structures formed during peat accumulation. Fyfe et al. (1983) noted evidence for bacterial sulfate reduction, with pyrite replacement of wood in the lignites. Pyrite-infilled bogen structure in fusinite (the characteristic open pores in charcoal) in the Mattagami Formation may also represent syndepositional or very early diagenetic phases, given similar observations in modern salt marsh sediments (White et al., 1990). Wilkin and Barnes (1997) demonstrated that framboids form in the water column of the Pettaquamscutt River estuary, and subsequently settle to the sediment-water interface, so the presence of framboids further suggests that much of the pyrite in the Mattagami Formation formed during deposition.

Carruccio et al. (1977) reported a direct correlation between overlying marine strata and the presence of framboids in coal, and White et al. (1990) found that extant marsh chemistries have an overprinting effect on underlying sediments producing relatively high amounts of pyrite beneath a marsh. Much of the pyrite contained within the Mattagami Formation paleosol thin sections exists as concentrically zoned inclusions bound within sphaerosiderites, indicating an overlapping paragenesis between the minerals. Sphaerosiderites represent mineralization during soil formation (Ludvigson et al., 1998), and siderite will not form in the presence of even minute quantities of H_2S (Berner, 1981). We suggest that pyrite unbound by sphaerosiderites in the paleosols may have formed through an overprinting process. This mode of formation is most applicable to higher %FeS_2 contents in some paleosols (particularly the lower paleosol in OGS 84-08; Fig. 3).

Source of Sulfate for Pyrite Formation in the Mattagami Formation

The most plausible source for sulfate is seawater. Modern dissolved sulfate concentrations for average marine waters are ~900 ppm (Brewer, 1975; Bruland, 1983), whereas the average total sulfur content of the major rivers of the world averages ~31 ppm (Martin and Meyback, 1979). Furthermore, low dissolved sulfide concentrations are commonly considered characteristic of anoxic, freshwater lake sediments (Hamilton-Taylor and Davison, 1995). These generalizations alone are suggestive of a marine sulfate source for pyrite formation in the Mattagami Formation. Nevertheless, other sources of dissolved sulfate are known to have contributed to pyrite formation in freshwater settings. Elevated total and pyritic sulfur values have been reported from ancient and modern evaporative saline lake systems (Tuttle et al., 1990). However, the intimate stratigraphic and paleogeographic relation of the Mattagami Formation laminated mudstones to kaolinitic mudrock paleosols and thick lignites suggests that evaporative conditions did not occur during formation of these deposits. In addition, the paleolatitudinal setting of the Moose River basin (~41° paleonorth) falls in the high precipitation belt indicated by climate simulations of the Albian–Cenomanian greenhouse world (Barron and Washington, 1985).

An upland source of evaporites could provide a source of dissolved sulfate to a drainage basin. For example, at Green Lake near Fayetteville, New York, dense, sulfate-rich bottom waters derived from groundwater flow through the gypsum-rich Silurian Syracuse and Vernon Shales lead to permanent stratification of the lake (Takahashi et al., 1968; Thompson et al., 1990). In this setting, pyrite is known to be forming in the water column and in the sediments (Suits and Wikim, 1998). The Moose River basin is known to contain evaporitic strata, but the vast majority of it crops out downgradient of paleoflow reconstructions of the Cretaceous Esoom River (map 1 of Telford, 1991). Therefore, it is unlikely that these strata contributed to a dissolved sulfate load for pyrite formation in the Mattagami Formation. Furthermore, the chemical composition of modern rainfall-dominated tropical rivers is controlled primarily by atmospheric composition (Berner and Berner, 1987), so any dissolved sulfate entering the Esoom River from the sliver of evaporitic strata which currently exists up the paleoflow gradient of the river was likely diluted to low concentrations. We concede that this sliver of strata might have provided a sulfate source, but pyrite excursions to concentrations >0.5% were more likely formed in a marine-influenced setting.

Stratigraphic and Environmental Implications of the Mattagami Formation

The pyrite-containing strata in the Mattagami Formation, along with the dinoflagellate-bearing laminated mudstones, are all near the palynologic zone 4–5 boundary of Norris and Zippi (1991). Palynologic zone 6 was proposed as late Albian; zone 5 was proposed to be earliest middle Albian to early late Albian, and zone 4 was proposed as early Albian (Norris and Zippi, 1991). We dispute these age assignments for the following reasons. First, most of the Mattagami Formation palynoflora share taxa with the upper Albian Dakota Formation in Iowa and Nebraska (Ravn and Witzke, 1995; Witzke et al., 1996; Witzke and Ludvigson, 1994). Second, although Norris and Zippi (1991, p. 119) assigned an early Albian–Aptian age to their zone 2 (their oldest zone), they stated that ''Zone 2 contains the basal range of *Liliacidites crassatus* which is confined to the upper Albian of Alberta.'' Furthermore, in an earlier study Norris (1982) rejected an Aptian–early Albian age assignment, and concluded that the deposits were middle and upper Albian. Third, key angiosperm taxa, which Norris and Zippi (1991) used to indicate an early to middle Albian age for their zone 4, all range at least into the late Albian (Zippi, 1992). In addition, angiosperm range zones are known to be diachronous across paleolatitudes in the Cretaceous (Norris, 1982), making their chronostratigraphic usefulness suspect. Norris and Zippi (1991) stated that zone 5 palynofloras are comparable to Joli Fou and Viking marine unit assemblages of the Canadian plains. The Joli Fou Formation is coeval to the Kiowa and Skull Creek Shales in the United States, whereas the overlying Viking Formation is coeval to the Muddy Sandstone (Vuke, 1984). The Kiowa–Skull Creek cycle was deposited during sea-level rise in the early late Albian (Kauffman and Caldwell, 1993), and ended the isolation of northern and southern arms of the Cretaceous Western Interior Seaway (Kauffman, 1977). This seaway confluence ended in the middle to late late Albian with a sea-level fall and development of a sequence boundary (Kauffman and Caldwell, 1993). The Muddy Sandstone has been interpreted as being deposited above the middle to late late Albian sequence boundary, and between the Kiowa–Skull Creek and overlying Greenhorn cycle (Weimer et al., 1988). Therefore, if the Mattagami Formation zone 5 palynoflora are coeval to the Joli Fou and Viking palynoflora, then the Mattagami Formation zone 5 must represent late Albian deposition.

Stacked channel sandstones and paleosols flanked by organic-rich clays and lignites (Fig. 7) are typical of the Mattagami Formation, and were interpreted as deposited near low-gradient anastomosed rivers in which bank stability was maintained by dense vegetation (Long, 1995). Stacking of fluvial sediments associated with thick lignites has been attributed to raised mires (McCabe, 1984; Warwick and Stanton, 1988). In this setting, avulsion may be inhibited because peat may be raised above river flood levels, and little sediment deposition occurs along these river segments (McCabe and Shanley, 1992). The thick Mattagami Formation lignites may have been formed as raised mires that deprived the Esoom River of a broad flood plain, and kept pace with the rate of base-level rise, a situation that McCabe and Shanley (1992) argued may have reduced the extent of Cretaceous Western Interior Seaway transgression. We suggest that the Mattagami Formation formed as a middle transgressive systems tract. In this setting, overbank flows, ponding, and flood-plain aggradation are widespread, and marshes colonize estuarine banks (Bohacs and Suter, 1997).

The facies associations between pyrite-bearing lignites, gray clays, and fluvial sandstones in the Mattagami Formation are interpreted here as having been deposited in an estuary. One definition of an estuary states that it is ''the area at a river mouth where salinities range from approximately 0.1 to 30–35‰'' (Dalrymple et al., 1992, p. 1130). Estuaries can be divided into three zones: (1) an outer marine-dominated zone; (2) a relatively low energy central basin zone; and (3) an inner river-dominated zone. The central basin typically contains the finest bedload sediment in the estuary, and acts as the prodelta region for the inner river-dominated zone if an open-water lagoon exists, or contains salt marshes and tidal channels in shallower estuaries (Dalrymple et al., 1992). Figure 7 includes a map view of an idealized estuary with our interpreted line of section for the Moose River basin cross section. We suggest that the cross section represents an oblique line from an easterly river-dominated portion of the Esoom River estuary, to a portion of the central basin in the

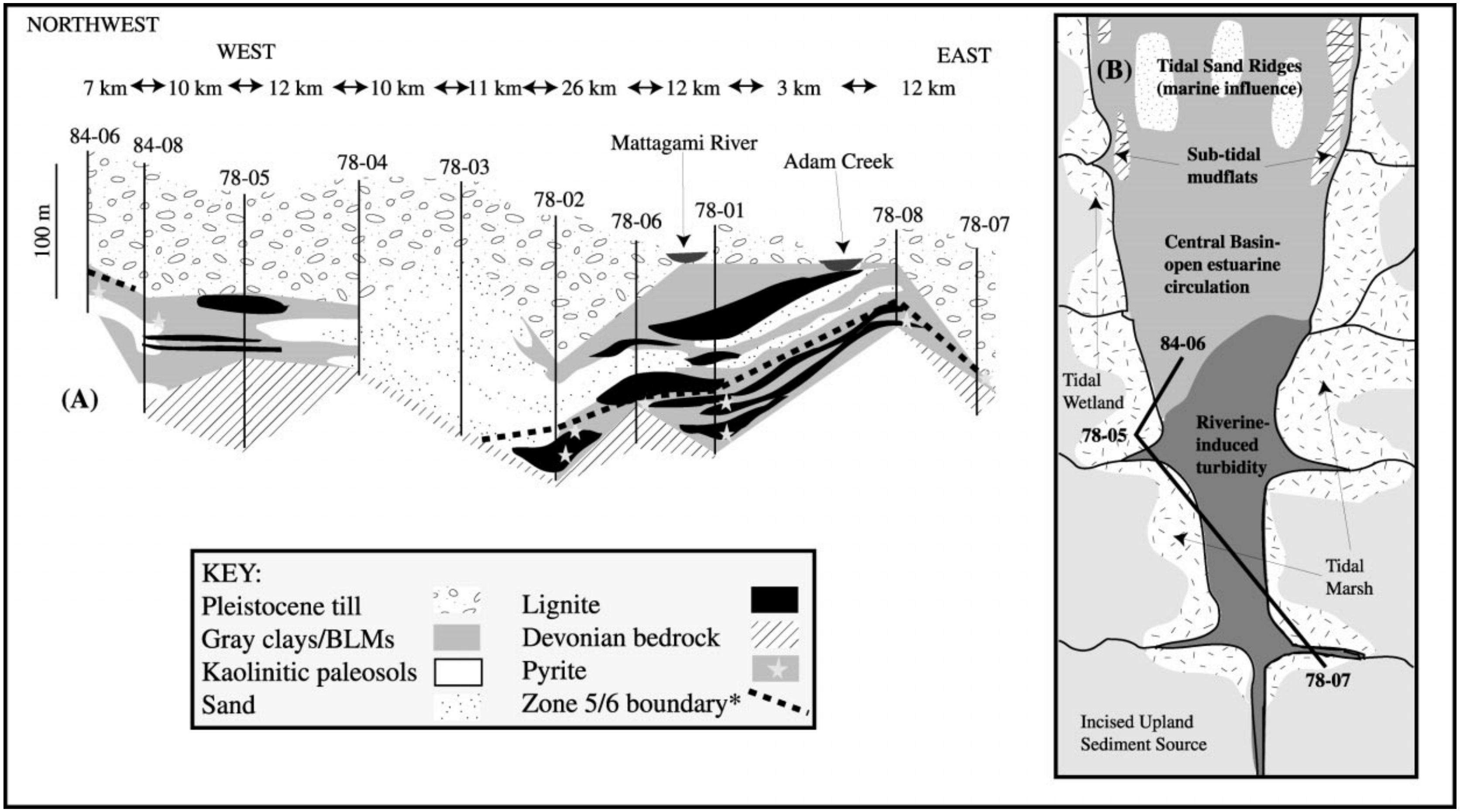

Figure 7. (A) Schematic east-west cross section in the Moose River basin showing lithofacies distribution modified from Norris and Dobell (1980) using lithologic core descriptions from Long (1991) and this study, and reported pyrite occurrences, and palynological zonation for the early to middle late Albian described in the text of this study. (B) Idealized representation of an estuary (modified from Dalrymple et al., 1992) showing our interpretation for placement of the Moose River basin cross section within this idealized scheme. Note that the key presents information only relevant to the cross section (A), and B is oriented such that the top of the figure is to the northwest.

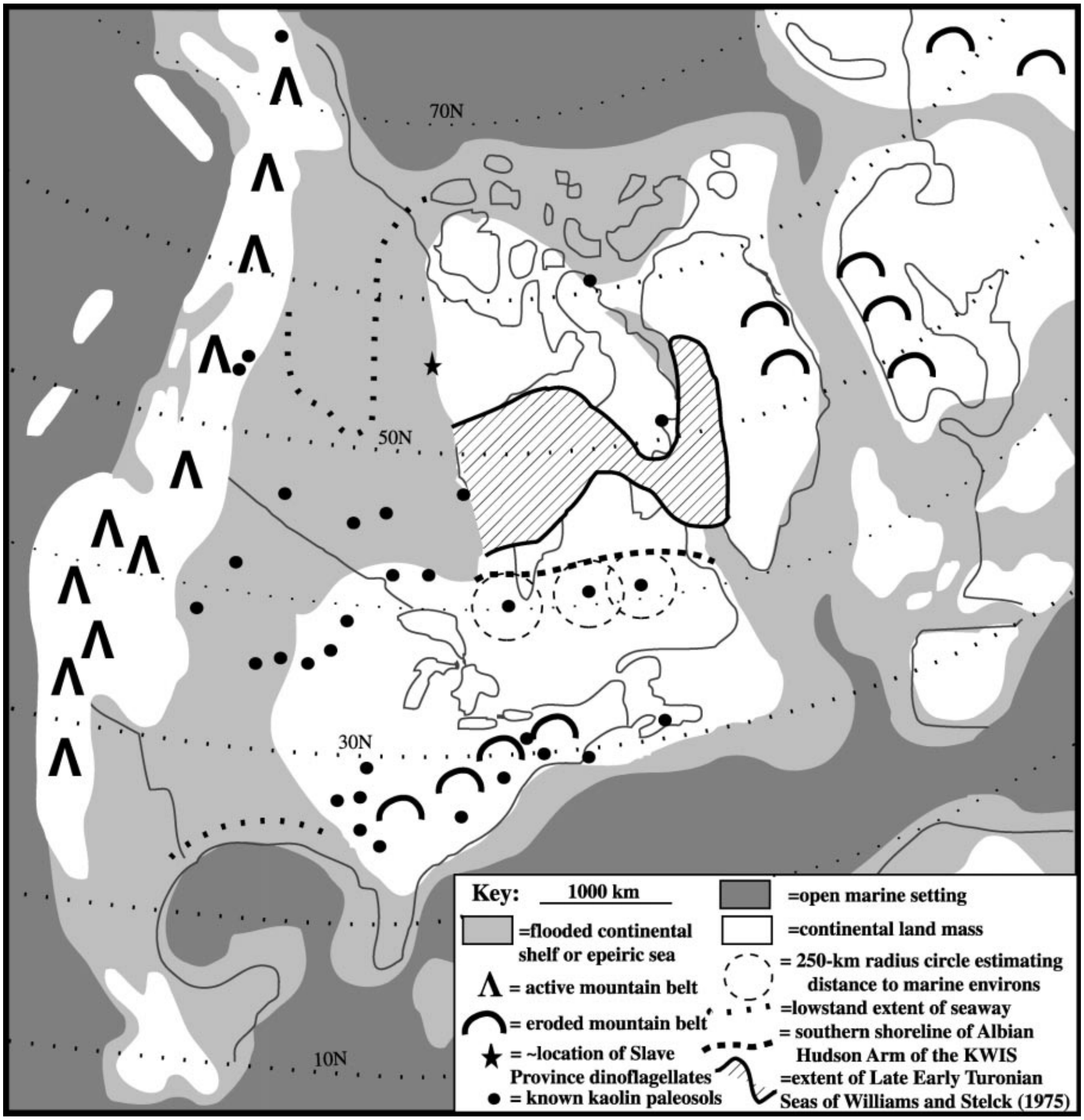

Figure 8. Middle to late Albian highstand paleogeography of North America (Kiowa–Skull Creek cycle; 98–106 Ma) showing the locations of known sphaerosiderite-bearing kaolinitic mudrock paleosols, in relation to our estimate for the southern paleoshoreline of a Hudson Arm connecting the proto–Labrador Sea to the Cretaceous Western Interior Seaway (KWIS). Note the extent of a late early Turonian Hudson Arm as determined by Williams and Stelck (1975). Also note the lowstand extent of the Western Interior Seaway to explain those paleosol locales plotted within the seaway, i.e., the shaded portions represent seaway highstand. Base map is adapted from Witzke and Ludvigson (1994), with modifications based on all the references cited in Figure 1, and Nassichuk and McIntyre (1995), Pindell and Barrett (1990), Tucholke and McCoy (1986), and Witzke and Ludvigson (1996) Paleolatitudes are interpolated from Scotese et al. (1988).

west. The thick lignites were deposited in an upstream position with interbedded channel sandstones likely representing a tributary to the main river. The main channel complex is shown in the middle of the cross section (i.e., cores 78-02 to 78-04). On the left side of the section, thinner less extensive lignites formed on the opposite side of the channel in a downstream position, and are associated with dinoflagellate-bearing laminated mudstones. We suggest that the dinoflagellate-bearing laminated mudstones were deposited in the central basin of the Esoom River estuary. This is consistent with paleocurrent-based reconstructions of the Esoom River (Fig. 8).

Pyrite in the Mattagami Formation is limited to the oldest strata in the formation (Fig. 7), which contains calcareous, medium quartz sand with lignite and authigenic pyrite; the younger phase contains fine quartz sand with kaolin (Hamblin, 1982). We suggest that the older phase represents initial marine flooding of the estuary. As the rate of relative sea-level rise slowed, the estuary backfilled with its fluvial bedload, and deposition of the younger phase occurred.

Late Albian dinoflagellate-bearing estuarine fills of the Dakota Formation, interfingering basinward to the Kiowa Shale, have been described in Iowa and Nebraska (Witzke et al., 1996). These deposits are considered marine influenced based on the presence of inclined heterolithic strata, marine palynomorphs (dinoflagellates and acritarchs), thalassinoid burrows, pyrite, and tidal rhythmites (Zawistoski et al., 1996). The significance of these estuarine fills coeval to the Mattagami Formation is in the occurrence of the dinoflagellate genus *Nyktericysta*, first named from the Kiowa Shale (Bint, 1986). We suggest that the widespread occurrence of this genus during the late Albian in North America is most easily described by dispersal through epeiric seas. We argue that the pyrite and dinoflagellate-bearing laminated mudstones of the Mattagami Formation were deposited during the early late Albian eustatic sea-level rise, and provide evidence for marine influence in the Moose River basin, analogous to strata of the Dakota Formation.

Kaolinitic Mudrock Paleosols and Implications to Paleogeography

The interfingering between dinoflagellate-bearing laminated mudrocks and kaolinitic paleosols is critical to our conclusion that the Mattagami Formation contains evidence of an Albian marine connection between the Labrador Sea and Cretaceous Western Interior Seaway through Hudson Bay. Table 1 is a compilation of 21 Albian–Turonian kaolinitic paleosols in North America, and their relationship to marine-influenced strata. More than half of the paleosol occurrences are considered to have formed proximally, i.e., <100 km, to marine water, and 20 of the 21 occurrences are within 250 km of the interpreted paleoshorelines. These locations are included on an Albian paleogeographic reconstruction of North America (Fig. 8), and 250-km-radius circles surrounding these sites approximate the nearest marine-influenced strata. We have demonstrated in this paper that the Mattagami Formation in the Moose River basin was marine influenced, so we consider the 250 km distance to be conservative. By confining the paleosols to a near coastal setting, we have estimated the position of a southern shoreline for the Hudson Arm of the Cretaceous Western Interior Seaway (Fig. 8).

Of particular interest to our reconstruction are red and gray clayey silts, siltstones, and sandstones with hematite and lignite in the Rupert Bay–Missisicabi River area in Quebec, described as lateral equivalents of the Mattagami Formation (Remick et al., 1963). Farther east, near the Quebec-Newfoundland border, the Cretaceous Redmond Formation contains clay, iron-rich clay, and iron concretions (Blais, 1959; Harrison et al., 1972), and carbonaceous material and brown coal as thick as 47 m (Usher, 1954). Blais (1959) indicated that these deposits are the equivalent of the Dakota Formation of the United States, and underlying clays are interpreted as Albian in age (Dorf, 1967). Some Redmond Formation concretions were reported to contain martite (Blais, 1959). Fernandez et al. (1990) reported that martite formed in soils as an intermediate in the oxidation of pyrite to hematite. Therefore, the Cretaceous Redmond Formation may represent coeval deposition to the Mattagami Formation, and may contain evidence for marine influence.

Structural Tectonic Setting and Basin Formation

Hamblin (1982) attributed preservation of Cretaceous sediments in the Moose River basin to faulting, whereas Blais (1959) noted overturned folds and thrust faults associated with deposition of the Cretaceous Redmond Formation. Harrison et al. (1972) noted normal faults and grabens associated with the Redmond Formation, and Gastil et al. (1960) hypothesized that the Redmond Formation iron-rich talus was deposited in a newly formed Cretaceous structural basin. Widespread igneous activity and regional faulting in eastern Canada and the northeastern United States have been associated with extension and opening of the Atlantic Ocean (Norris, 1993). However, Faure et al. (1996) attributed the regional structural fabric of eastern Canada and New England to the extensional regime associated with the initiation of rifting between Labrador and Greenland ca. 140 Ma. Cox and Van Arsdale (1997) suggested that the St. Lawrence rift system formed by passage of the Great Meteor hotspot beneath the region in the Cretaceous (Cox and Van Arsdale, 1997). Their model includes uplift and erosion when the hotspot was situated beneath the region, and subsidence associated with the passage of the hotspot away from the area (Cox and Van Arsdale, 1997). Note that the easternmost extent of our interpreted Hudson Arm southern paleoshoreline (Fig. 8) intersects the proto–Labrador Sea, and that Albian marine strata are known to exist in the Labrador Sea (Burden and Langille, 1991).

Crough et al. (1980) explained the progressively older-to-the-west ages of the Cretaceous New England Seamounts, and their alignment with the Cretaceous intrusions of New England and eastern Canada, by the westward migration of North America over the Great Meteor hotspot. Furthermore, Crough (1981) suggested that maximum uplift and erosion occurred along the Great Meteor hotspot track with less uplift and erosion toward the margins. Duncan (1984) showed predicted hotspot tracks and the age of igneous intrusions associated with the New England and Newfoundland Seamounts. The most southerly of these tracks extends through the Moose River basin, while a second track traces through the Rupert Bay–Missisicabi River area in Quebec. If the ages of igneous intrusions presented by Duncan (1984) are correct, then the hotspots were located offshore eastern North America by the late Albian.

The extensional and hotspot models provide mechanisms for accommodation space development. Regional extension may have weakened the crust, prior to the passage of eastern North America over at least one hotspot. Fault (re)activation, uplift, erosion, and subsidence likely occurred at this time. Note that the northern extent of the predicted hotspot tracks (Fig. 1) coincides with our southern shoreline for the Albian Hudson Arm of the Cretaceous Western Interior Seaway (Fig. 8). We suggest that the Albian shoreline may have formed along a northern deformation front. The existence of an Albian Hudson Arm is supportive of a Cenomanian–Turonian marine connection previously proposed across the region, since eustatic sea level is considered to have been higher during the Cenomanian–Turonian highstand (Haq et al., 1988). The presence of a Hudson Arm seaway likely played an important role in climate moderation, ocean circulation in the Cretaceous Western Interior Seaway, and floral and faunal dispersal patterns, and therefore should be considered in any paleoenvironmental reconstructions of Cretaceous North America.

CONCLUSIONS

Dinoflagellate-bearing laminated mudstones of the Albian Mattagami Formation, Moose River basin, James Bay Lowlands, Ontario, previously interpreted as freshwater in nature, contain a record of marine influence which we attribute to the existence of a Hudson Arm between the Labrador Sea and the Cretaceous Western Interior Seaway of North America. This reinterpretation is based on mutually consistent relationships between total iron, pyritic sulfur, and organic matter

TABLE 1. STRATIGRAPHIC COMPILATION OF KNOWN ALBIAN–TURONIAN PALEOSOLS IN NORTH AMERICA AND THEIR RELATIONSHIP TO THE NEAREST MARINE STRATA

Location	Age	Stratigraphic separation from marine units (m)	Paleogeographic distance to coeval marine strata (km)	References
Iowa				
Sioux City	Albian–Cenomanian	30–100	180–250	Ravn and Witzke (1995); Ludvigson and Witzke (1996); Witzke et al. (1996)
Nebraska				
Jefferson County	Albian–Cenomanian	10	Proximal–120	Witzke et al. (1996); R. Ravn (1998, personal commun.); Ludvigson and Witzke (1996)
Kansas				
Jones core	Albian–Cenomanian	Intercalated	Proximal	Witzke et al. (1996); Ludvigson and Witzke (1996); R. Ravn (1998, personal commun.)
Bounds core	Albian–Cenomanian	Intercalated	Proximal	Scott et al. (1998)
Minnesota				
Minnesota River valley	Cenomanian	10	Proximal–220	Witzke et al. (1996); R. Ravn (1998, personal commun.); T. White (measured section)
Ontario				
Rainy River core	Late Albian	(?) 3 reworked dinoflagellates	180	Zippi and Bajc (1990); Williams and Stelck (1975)
Moose River basin	Late Albian	Intercalated	Proximal	Palynology cited in text; this study
Manitoba				
Swan River	Late Albian	~15	Proximal	Williams and Stelck (1975); McNeil and Caldwell (1981); R. Ravn (1998, personal commun.)
Saskatchewan				
Yarbo core	Late Albian	~15	Proximal	Williams and Stelck (1975); White and Witzke (measured core section); R. Ravn (1998, personal commun.)
Alberta				
Bow Island Formation	Middle Albian	Intercalated transgressive surfaces	100–150	McCarthy et al. (1997a, 1997b)
British Columbia				
Boulder Creek Formation	Late Albian	Intercalated	Proximal	Leckie et al. (1989)
Dunvegan Formation	Cenomanian	5	50–200	McCarthy and Plint (1998)
Wyoming				
Powder River basin	Albian–Cenomanian	Immediately overlain	<100	Weimer et al. (1988); Elder and Kirkland (1994)
Colorado				
Denver basin	Albian–Cenomanian	Immediately overlain	<100	Weimer and Sonnenberg (1989); Elder and Kirkland (1994)
Utah				
Henrieville	Middle Cenomanian	20	150	Kirschbaum and McCabe (1992); Elder and Kirkland (1994)
Kaiparowits Plateau	Middle to late Turonian	10	100	Elder and Kirkland (1994); Hettinger (1995)
New Jersey				
Patapsco Formation	Albian–Cenomanian	30	<200	Owens et al. (1977); Pete Sugarman (New Jersey Geological Survey, 1998, personal commun.)
Raritan Formation	Cenomanian–Turonian	Immediately overlain	<100	Pete Sugarman (New Jersey Geological Survey, 1998, personal commun.); T. White (core description)
Maryland/Delaware				
Potomac Group	Albian–Turonian	Immediately overlain	<100	Glaser (1969); Pickett (1987); K. Ramsey (Delaware Geological Survey, 1998, personal commun.)
Georgia/Alabama				
Tuscaloosa Group	Cenomanian	<50	~120	Gohn (1988); Sigleo and Reinhardt (1988); Sohl et al. (1991)
Alaska				
Nanushuk Group, North Slope	Albian–Cenomanian	intertongueing shallow marine shale and sand	Proximal–250	Ahlbrandt et al. (1979); Huffman (1985); Moore et al. (1994)

in the sediments. Relatively high quantities of labile organic matter and pyrite are associated with horizons within the mudstones that are reported to contain dinoflagellates. The dinoflagellate genus has been described elsewhere in coeval marine-influenced strata. We conclude that the Mattagami Formation mudstones were deposited in an estuarine system that flowed into the Hudson Arm. Our reconstruction of the southern paleoshoreline of the Hudson Arm shows that the marine connection to the Labrador Sea developed through a region subject to Cretaceous normal faulting. We interpret this relationship as indicating that the Hudson Arm basin may have been related to the extensional regime associated with initial rifting of the Labrador Sea basin, and/or the passage of North America over the Great Meteor hotspot during the Cretaceous.

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