PALEOCEANOGRAPHY OF THE SOUTHWESTERN WESTERN INTERIOR SEA DURING THE TIME OF THE CENOMANIAN-TURONIAN BOUNDARY (LATE CRETACEOUS)

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ABSTRACT: The Cenomanian-Turonian boundary interval (93-94 Ma) was a time of rapid oceanographic change in the U.S. Western Interior Sea ("Greenhorn Sea"). Previous studies documented changes in $\delta^{18}O$ in carbonates and shifts in macrofossil (molluscan) populations indicating the incursion of a subsaline surface-water mass into the region and dysoxic to anoxic benthic conditions across wide areas of the seaway. These changes were accompanied by an expanding oxygen minimum zone in concert with global-scale burial of organic matter, which was driven in part by elevated rates of marine productivity. The oceanography of the southern seaway was undoubtedly complex with signals of a global anoxic event overprinted by regional influences of water mass stratification, mixing, productivity, changes in relative sea level, and biotic turnover.

Our studies of planktic and benthic foraminiferal assemblages and clay-mineral distribution in calcareous mudrocks and dark marlstones from the southwestern side of the Greenhorn Sea provide further constraints on the paleoceanography of this region. Specifically, there is a major change in planktic foraminiferal population structure ("*Heterohelix* shift") coupled with an influx of kaolinite and illite at both a neritic site (Lohali Point, Arizona) and a distal basin site (Rock Canyon, Colorado). These characteristics are not observed at a proximal basin site (Mesa Verde, Colorado) in between the other locales. The distinctive clay mineral assemblages suggest two disparate sources. One of these sources most likely was from the Sevier orogenic belt along the western side of the seaway, and another was either the southwestern corner of the seaway or the southern part of the stable craton. The distribution of clay mineral and planktic foraminiferal assemblages provide information on circulation of the upper water column.

The benthic foraminiferal assemblages of Lohali Point and Mesa Verde are very similar and have northern affinities, suggesting the influence of cool bottom waters along the western side of the seaway. We suggest that a submerged tectonic forebulge or bathymetric high near Mesa Verde caused "edge-effect" mixing and upwelling of cool water masses originating from the north. To the west, a foredeep shelf in northeastern Arizona and south-central Utah provided a conduit for northward-flowing, warmer surface water masses over the southward-flowing, cooler waters. These southern waters were bifurcated by the Mesa Verde high. The resultant oceanographic front, or mixing zone, caused the contrast in ecological and sedimentological patterns at the sites. With rising sea level came the incursion of oxygen-poor Tethyan intermediate waters into the Greenhorn Sea during latest Cenomanian-early Turonian time and the development of widespread, low diversity benthic foraminiferal assemblages dominated by *Neobulimina*.

INTRODUCTION

The Cretaceous Western Interior Sea had a long and dynamic oceanographic history. Its marine waters were strongly influenced by northern water masses for much of this history. The paucity of many normal marine invertebrates, such as echinoderms, bryozoans, corals, and rudist bivalves, suggests that the salinity of the sea was lower than normal (Kauffman, 1975, 1977, 1984). The Western Interior Sea received runoff from numerous rivers that drained highlands on the western side of the seaway and lowlands of the stable craton on the eastern side. During episodes of high eustatic sea level and transgression, subtropical normal marine water masses invaded the seaway from the Tethys Sea to the south. The warm water masses penetrated as far north as the prairie provinces of southern Canada (Kauffman, 1984; Caldwell et al., 1993; Kauffman and Caldwell, 1993). With these transgressions came a more normal marine biota, mixing and juxtaposition of different water masses, and changes in regional climate and sedimentation patterns. One such episode occurred during the late Cenomanian and early Turonian when the seaway, also called the "Greenhorn Sea", expanded to nearly 2000 km wide during the highest eustatic sea level rise of the Mesozoic Era (Hancock and Kauffman, 1979; Haq et al., 1988; Hay et al., 1993).

The axis, or deepest part of the seaway was probably along the corridor from northeastern New Mexico to the Black Hills region of northeastern Wyoming and southwestern South Dakota (e.g., Eicher, 1969b; Kauffman, 1977; Sageman and Arthur, 1994) (Fig. 1). This is where the salinity-sensitive planktic foraminifera first invaded the hyposaline interior sea when sea level rose during the late Cenomanian and where they persisted the longest with the subsequent fall of eustatic sea level during the middle Turonian (Eicher, 1969a; Eicher and Worstell, 1970; Eicher and Diner, 1985). It is also the corridor through which benthic foraminifera with southern affinities invaded the seaway.

The rapid incursion of subtropical water masses into the seaway during late Cenomanian time abruptly improved benthic conditions and brought an invasion of warm water molluscs, foraminifera, and calcareous nannofossils (Cobban and Reeside, 1952; Reeside, 1957; Eicher and Worstell, 1970; Kauffman, 1977; McNeil and Caldwell, 1981; Eicher and Diner, 1985; Elder, 1985, 1991; Elder and Kirkland, 1985; Bralower, 1988; Caldwell et al., 1993; Hay et al., 1993; Kauffman and Caldwell, 1993; Watkins et al., 1993; Fisher et al., 1994). This event also coincided with the onset of a global perturbation in the carbon system, referred to as an Oceanic Anoxic Event (OAE-2), which was related in part to a brief interval of elevated marine productivity and burial of organic carbon (Arthur et al., 1987, 1988, 1990; Schlanger et al., 1987). The carbon event, measured in terms of $\delta^{13}C$ of carbonates and organic matter, was probably intimately linked with tectonics, eustatic sea level rise, oceanic circulation, and mode of deep water mass formation.

The global OAE-2 had a profound effect on some of the marine plankton, nekton, and benthon of the Greenhorn Sea. For example, Elder (1991) delineated multiple extinction steps in western interior sections through the uppermost Cenomanian and lower Turonian (Fig. 2), and Eicher (1969a) and Leckie (1985) documented a major turnover in planktic foraminifera through the Cenomanian-Turonian boundary interval. Therefore, the Western Interior Sea responded to major global changes that were modified by regional climatic, oceanographic, and deposi-

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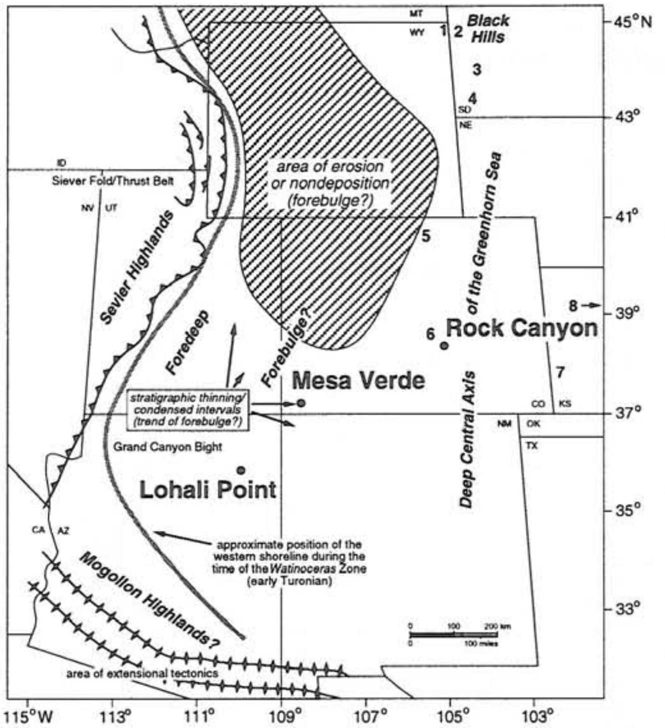

FIG. 1— Base map showing the southwestern portion of the Cenomanian-Turonian Western Interior Sea ("Greenhorn Sea"). Map is centered over Wyoming, Utah, Colorado, Arizona, and New Mexico. Locations of the three primary sites discussed in the text are: Lohali Point (LP), Black Mesa Basin, northeastern Arizona, a neritic site; Mesa Verde (MV), northern San Juan Basin, southwestern Colorado, a proximal basin site; Rock Canyon (RC), near Pueblo, Colorado, a distal basin site. Numbered localities show sections of Eicher and Worstell (1970) Section 8 is located in central Kansas. Map information from McGookey et al. (1972), Molenaar (1983), Cobban and Hook (1984), Kauffman (1984, 1985), Hattin (1985, 1987), Merewether and Cobban (1986), Eaton et al. (1990), Eaton and Nations (1991), Kirkland (1991),

tional signals. A major focus of this paper is to examine the record of regional oceanographic change along the southwestern side of the Greenhorn Sea, as documented by clay mineralogy and foraminiferal paleoecology and paleobiogeography. Together with earlier studies of stable isotopes and molluscan paleoecology and paleobiogeography, we evaluate the impact of Cretaceous global change on this large epicontinental seaway.

LOCATION AND METHODS

Measured Sections

Three sections through the Cenomanian-Turonian stage boundary were examined, that represent a transect across the southwestern side of the Cenomanian-Turonian Greenhorn Sea (Fig. 1). According to Elder's (1991) lithofacies regions, Lohali Point (LP) is in the clay-dominated western region, Mesa Verde (MV) is in the transitional lithofacies region, and Rock Canyon (RC) is in the carbonate-dominated central axis of the Greenhorn Sea. There is a progressive change in the siliciclastic and carbonate content of the strata from the western margin to the basin center (Hattin, 1985, 1986b) which reflects proximity to the tectonically active Sevier orogenic belt on the west. In general, sediment accumulation rates were two to four times greater in the western, clay-dominated lithofacies than in the transitional and carbonate-dominated lithofacies (Elder, 1985; Elder and Kirkland, 1985). The carbonate-rich sections are characterized by interbedded, light-colored limestones and dark marlstones or calcareous shales of the Bridge Creek Member through the Cenomanian-Turonian boundary interval (Mancos Shale at Mesa Verde; Greenhorn Formation at Rock Canyon). Our samples come from the muddy and marly intervals only; no limestones were investigated.

Lohali Point (LP), located on the eastern side of the Black Mesa Basin in northeastern Arizona, represents a neritic depositional environment. This area was part of a north-south trending "foredeep" that developed seaward of the Sevier orogenic belt. The foredeep contained the thickest accumulation of sediments shed off the adjacent highlands, but this region probably did not represent the greatest water depths (Kauffman, 1977, 1984; Kirkland, 1991). LP is near the distal end of a broad, shallow, seaward-sloping shelf that covered northern Arizona and south-central Utah ("Grand Canyon Bight") during latest Cenomanian-middle Turonian time (Eaton and Nations, 1991; Kirkland, 1991; Leckie et al., 1991; Leithold, 1993). The lithostratigraphy and ammonite biostratigraphy of the lower shale member of the Mancos Shale are from Elder (1987) and Kirkland (1991).

The Mesa Verde (MV) section represents a proximal basin setting located at the northern end of Mesa Verde National Park in southwestern Colorado. There is evidence for stratigraphic thinning and condensed intervals in the upper Graneros and lower Bridge Creek Members of the Mancos Shale (uppermost Cenomanian-lower Turonian), similar to the Red Wash section south of MV in the northwestern corner of New Mexico (Elder, 1985, 1991; Leckie et al., 1997). The four-corners area may have been along the trend of a bathymetric high, or forebulge, at this time. The stratigraphic framework for this study comes from Kirkland et al. (1995) and Leckie et al. (1997).

The Rock Canyon section (RC) includes the uppermost Hartland Shale and lower Bridge Creek Limestone Members of the Greenhorn Formation exposed west of Pueblo, Colorado. This represents a distal basin depositional setting from within the deepest part of the seaway (Eicher, 1969a, b; Cobban and Scott, 1972; Hattin, 1971, 1985, 1986b; Kauffman, 1977, 1984; Kauffman, Pratt, et al., 1985). The molluscan biostratigraphy and lithostratigraphy used in this study are from Elder (1985), Elder and Kirkland (1985), and Sageman (1985).

Foraminiferal Assemblage Analyses

We use planktic and benthic foraminiferal sediment assemblages to infer environmental changes in the upper water column and near the seafloor. The analyses focus on the following: (1) proportion of planktics to benthics in the total foraminiferal assemblage (percent planktics), (2) major groups of planktic genera or morphotypes (biserial *Heterohelix*, triserial *Guembelitria*, trochospiral *Hedbergella* and *Whiteinella*, planispiral *Globigerinelloides*, and the keeled genera *Rotalipora*, *Praeglobotruncana*, and *Dicarinella*), and (3) major groups or species of benthics (*Neobulimina albertensis*, *Gavelinella dakotensis*, other calcareous benthic taxa, and agglutinated taxa). Previous studies documented the utility of these taxa or morphogroups in Cenomanian-Turonian rocks from the western interior (e.g., Eicher and Diner, 1985; Leckie, 1985; Leckie et al., 1991; Fisher et al., 1994; Schroeder-Adams et al., 1996). The assemblages are based on counts of approximately 300 specimens from the >63µm size fraction. Many of the samples from the Rock Canyon section were picked a second time to increase

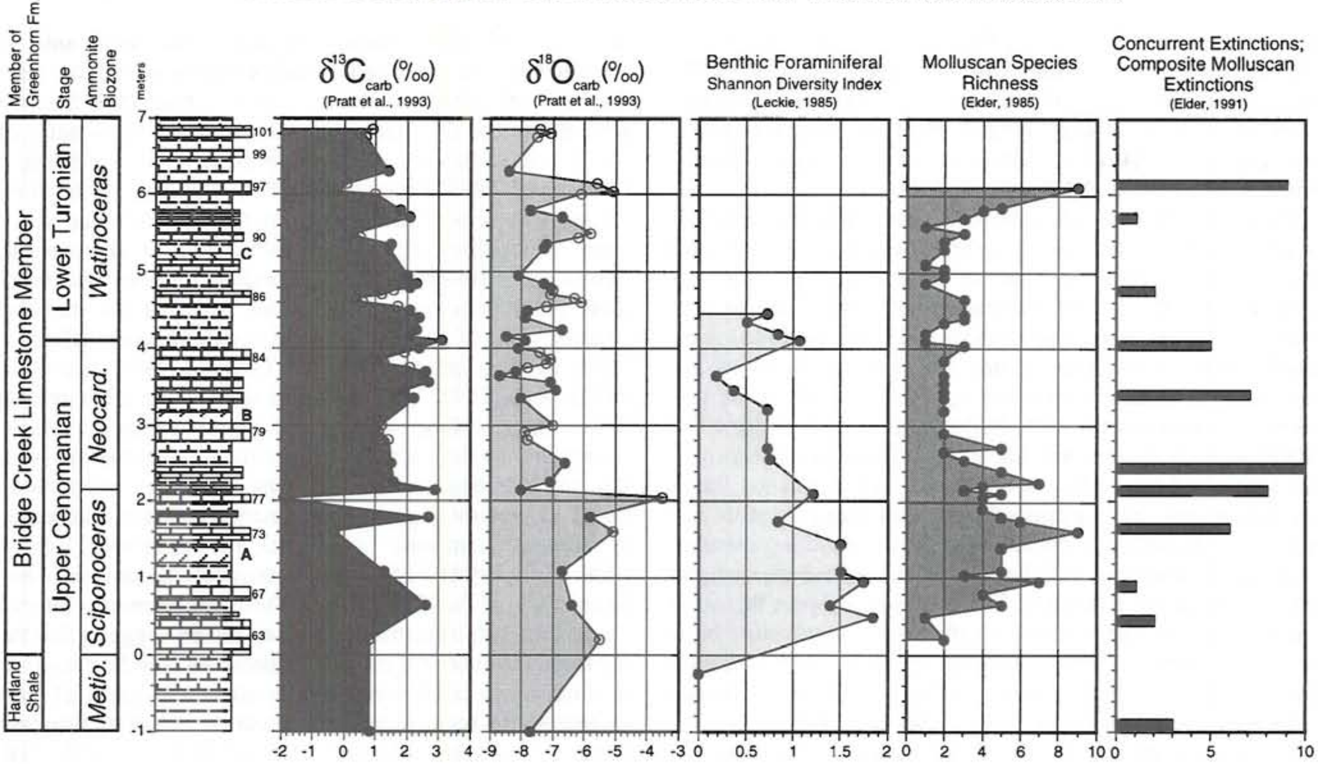

FIG. 2—Details of the Cenomanian-Turonian boundary interval at the Rock Canyon section (near Pueblo, Colorado). Ammonite biostratigraphy, lithostratigraphy, and bentonite beds (A-C) are from Elder (1985) and Elder and Kirkland (1985); limestone bed numbers are from Cobban and Scott (1972). Isotope data are from Pratt et al. (1993); open circles are limestones and closed circles are dark marlstones or calcareous shales. Benthic foraminiferal diversity data [Shannon-Wiener diversity index, H(S)] are from Leckie (1985). Molluscan extinction steps are from Elder (1991).

the benthic assemblages to at least 300 specimens (Leckie, 1985). All samples were also examined for rare species and other biogenic and mineral components. For details of sample processing, picking, and counting, refer to Leckie et al. (1991).

Clay Mineralogy

Our analyses focused on the 1-2µm size fraction, a bit of a departure from most investigations of clay-mineral abundance that examine the entire <2 or <1µm size. Because of the dominance of mixed-layer illite/smectite in the broad fractions, we examined the coarsest subset to enhance the relative contributions of other clays, which often have larger grains (Gibbs, 1977). The grain sizes were separated by centrifuging for the appropriate length of time. For details of sample preparation, see Leckie et al. (1991). Oriented mounts were prepared by smearing a clay paste on glass slides (Gibbs, 1971). X-ray diffraction of the samples was accomplished with a Siemens diffractometer updated with a Databox digital data processing system. Samples were scanned untreated and after solvation with ethylene glycol. Selected samples were also heated to 550°C or Mg-saturated and solvated with glycerol to further help identify the clay-mineral phases (Brindley and Brown, 1980; Hardy and Tucker, 1988).

Obtaining quantitative mineral abundance from X-ray diffractograms has always been a vexing problem. Although the techniques for sample preparation and analysis are being refined so that X-ray intensities can be converted more meaningfully into mineral percentages, the techniques are not always amenable to processing large numbers of samples (Moore and Reynolds, 1989). Simple measurements of relative peak height (or area) are adequate for semi-quantitative comparisons of similar samples (Yuretich, 1979; Leckie et al., 1991). This is especially true when relative changes are the critical component, as opposed to absolute amounts of a particular mineral in a sample. For these reasons, our clay studies focus on the most easily recognized and measurable X-ray diffraction peaks in the clays. Relative abundance of clay minerals was determined by ratios of diffractogram peak height of the 001 diffraction line of each clay (Griffin, 1971).

BACKGROUND

Tectonics

The Cretaceous, the mid- to Late Cretaceous in particular, was characterized by heightened plate tectonic activity, including elevated rates of seafloor spreading, convergent margin volcanism, and intraplate volcanism (e.g., Larson, 1991a, b; Arthur et al., 1985, 1990). More than 700 volcanic ashes (bentonites) are described from upper Albian to middle Maastrichtian strata of the U.S. western interior (Kauffman, 1985). The Sevier Orogeny refers to tectonic deformation during Cretaceous time that occurred along a narrow but continuous fold and thrust belt, extending from southern Nevada to the Canadian Rockies (Armstrong, 1968; McGookey et al., 1972). The tectonism was associated with large-scale magmatic activity and terrain accretion along the convergent western margin of the North American cordillera (e.g., Schwartz and DeCelles, 1988). Magmatism reached its greatest extent at about 90-95 Ma, roughly coincident with the peak of Cretaceous sea level (Armstrong and Ward, 1993). The eastward-vergent tectonic loading was responsible for creating a large, asymmetric foreland basin in the Canadian and U.S. western interior (e.g., Price, 1973; Cross and Pilger, 1978; Beaumont, 1981; Jordan, 1981; Wiltschko and Dorr, 1983; Schedl and Wiltschko, 1984). The axis of maximum sediment accumulation in the western interior foreland basin was immediately adjacent to the thrust belt, also called the foredeep (e.g., Lawton, 1985; Villien and Kligfield, 1986).

Within the foreland basin there is abundant sedimentologic, lithostratigraphic, and biostratigraphic evidence of common subtle intraforeland uplifts. Many of these structural highs were intermittent and short-lived features on the seafloor of the interior seaway. For example, Merewether and Cobban (1986) documented numerous elongate arches and mild swells of middle Cenomanian to middle Coniacian age, many of which are progressively younger to the east. Heller et al. (1993) suggest that such topographic features are due to modest changes in intraplate stress levels and are to be expected in a heterogeneous lithosphere with a long, but punctuated history of changing plate motion stresses. In addition, differential movement along preexisting planes of weakness may have accentuated the topographic relief of some of these features. There is also evidence that some of the Cretaceous intraforeland uplifts reflect reactivated basement uplifts, some of which are precursors to later Laramide deformation (e.g., Ryer and Lovekin, 1986; Hattin, 1987; Schwartz and DeCelles, 1988; Eaton et al., 1990; Heller et al., 1993).

Numerous authors have speculated about the existence of a forebulge adjacent to the axis of maximum subsidence and sediment accumulation (e.g., Lorenz, 1982; Kauffman, 1984, 1985; Merewether and Cobban, 1986; Eaton et al., 1990; Leithold, 1994; Christie-Blick and Driscoll, 1995). One possible candidate for a forebulge or topographic high during middle Cenomanian-early Turonian time is the arch or series of arches that developed across western Wyoming, northeastern Utah, and western Colorado (Merewether and Cobban, 1986; Ryer and Lovekin, 1986; Molenaar and Cobban, 1991; Franczyk et al., 1992). This area is delineated by a widespread lacuna (Elder and Kirkland, 1993 a, b; Sageman and Arthur, 1994). The broad area of uplift apparently migrated across central Wyoming, and central and northeastern Colorado during middle and late Turonian time (Merewether and Cobban, 1986).

Further to the south there is also evidence of stratigraphic thinning and erosion on presumed topographic highs during late Cenomanian-early Turonian time in eastern Utah along the present trend of the San Rafael Swell and Monument Uplift (Eaton et al., 1990; Elder, 1991; Leithold, 1994), in southwestern Colorado (Lamb, 1968; Elder, 1991; Leckie et al., 1997), and in easternmost Arizona along the present trend of the Defiance Uplift (Elder, 1991). Elder (1991) commented on the importance of this general north-south structural trend in controlling Cenomanian-Turonian lithofacies distribution on the Colorado Plateau, with the thicker, higher sedimentation-rate, clay-dominated sequences confined to the west (foredeep). Toward the basin center to the east, stratigraphic thinning in uppermost Cenomanian and basal Turonian strata of the Bridge Creek Member (Greenhorn Formation) delineate another probable bathymetric high trending northwest-southeast across south-central Colorado and northeast New Mexico (Hattin, 1987).

Eustasy and Relative Sea Level

The transgression and regression of the latest Albian-middle Turonian Greenhorn Sea represents a third-order, largely eustatic, rise and fall of sea level (Kauffman, 1977, 1984, 1985; Kauffman and Caldwell, 1993). Superimposed on this dominantly global sea level record are smaller scale changes in relative sea level that are controlled by a combination of factors including eustasy, episodic uplift and basin subsidence, and climatically driven changes in sediment supply (Kauffman, 1985; Leithold, 1993, 1994; Ryer, 1993; Elder et al., 1994). Relative sea level rise, as recorded in upper Cenomanian and lower Turonian sediments from the southwestern side of the seaway, is characterized by fourth- and fifth-order marine flooding surfaces and back-stepping, progradational sequences that comprise the transgressive systems tract of the third-order Greenhorn cycle (Elder et al., 1994; Gardner and Cross, 1994; Leithold, 1994; Christie-Blick and Driscoll, 1995; West et al., this volume). In offshore mudrock facies, the flooding surfaces and parasequence boundaries are recognized by thin intervals of sediment condensation (winnowing, calcisiltite/calcarenite horizons, shell lags, concretion horizons), a decrease in grain size, elevated levels of carbonate, and/or influx of warm water taxa (Elder, 1991; Leithold, 1993, 1994; Elder et al., 1994). Elder (1991) notes that the ammonite biozone boundaries of the Cenomanian-Turonian boundary interval are associated with transgressive pulses and he suggests that the faunal turnover observed at these boundaries may be closely linked to relative sea level events in the seaway in terms of changing sedimentation rates, benthic oxygenation, substrate firmness, and water temperature. Similarly, Li and Habib (1996) attributed changes observed in dinoflagellate diversity, assemblage composition, and organic facies to changes in sea level during late Cenomanian and earliest Turonian time.

Climate

The striking development of interbedded, light-colored limestones and dark marlstones or calcareous shales in the Bridge Creek Limestone Member of the Greenhorn Formation has long been attributed to climatic cyclicity (Gilbert, 1895; Kauffman, 1977; Fischer, 1980; Pratt, 1984; Barron et al., 1985; Fischer et al., 1985; Eicher and Diner, 1989). Individual limestone beds can be traced over 1000 km, which attests to the remarkable uniformity across the vast central part of the seaway (Hattin, 1971, 1985, 1986b). Periodicities in the Bridge Creek Member are believed to roughly correspond with Milankovitch climate forcing dominated by the 41 ka obliquity cycle (Kauffman, 1977; Fischer, 1980; Fischer et al., 1985).

Rhythmic bedding of the Bridge Creek Member is best developed in the deep, distal, central part of the seaway late in transgression, following incursion of warm water masses. This represents the time and place of heightened contrasts between water masses and west-east patterns of sedimentation. Enhanced limestone-shale cyclicity was due, in part, to decreased sedimentation rates with rising sea level and increased calcareous plankton productivity with the incursion of warm normal marine waters into the seaway with transgression (e.g., Elder, 1985, 1991; Hattin, 1986b). The marlstones and calcareous shales are darker in color due to higher relative concentrations of clay minerals and organic matter (Pratt, 1984). These lithologies are either laminated or only weakly bioturbated. The limestones are composed mostly of biogenic carbonate from calcareous nannofossils and planktic foraminifera (Hattin, 1971; 1986a). These latter beds are highly bioturbated and contain less organic matter and terrigenous material (Pratt, 1984; Barron et al., 1985).

According to the dilution model of Pratt (1984), the shalier or muddier lithologies of the Bridge Creek Member represent times of greater rainfall and runoff. Accordingly, episodes of high riverine discharge to the seaway result in the development of a sediment-laden, reduced-salinity surface layer, heightened terrigenous influx, poor benthic ventilation due to salinity stratification, and weak deep water circulation (Pratt, 1984, 1985; Barron et al., 1985; Hattin, 1985, 1986b; Bottjer et al., 1986; Pratt et al., 1993). The limestones, on the other hand, represent times of drier climate, reduced terrigenous influx, near-normal marine salinities from surface to deep, and improved circulation and benthic oxygenation. The darker marlier and shalier lithologies are commonly thicker than the adjacent limestone beds in the central part of the seaway (e.g., Hattin, 1985, 1987), which suggests higher rates of siliciclastic sedimentation during periods of higher precipitation and runoff. In addition, increasing burrow size, density, and depth of burrow penetration are correlated with decreasing organic carbon and increasing carbonate content (Savrda and Bottjer, 1993). This evidence suggests that climate-driven cycles of clastic dilution and benthic ventilation are the principal causes of rhythmic bedding in the Greenhorn Formation (Hattin, 1971; Pratt, 1984; Barron et al., 1985; Savrda and Bottjer, 1993; Pratt et al., 1993).

Clay minerals often indicate the weathering conditions in the source area from which they were derived and, in the absence of subsequent diagenetic alteration, the relative abundance of different kinds of clays can be used to indicate paleoclimate. Chamley (1989) and Singer (1980, 1984) summarized the relationships gathered from numerous studies of modern sediments and sedimentary rock sequences. In cold climates, physical weathering predominates and the fine-grained components of soils are generally fragments of parent minerals. Illite and chlorite are frequently the most abundant clay minerals, since the lithologies in many high-latitude areas are dominated by igneous and metamorphic bedrock. In more temperate climates, chlorite decomposes into vermiculite, and illite will hydrolyze to form smectitic clays. Under these climatic conditions, the abundance of smectite generally increases as the amount of annual rainfall increases. Under warm, tropical conditions, chemical breakdown of parent materials produces aluminous chemical residua with larger proportions of kaolinite and gibbsite. Again, the more aluminous end members result from the highest rates of leaching, which is generally a function of rainfall amount. These generalizations can be modified owing to the source lithology and drainage characteristics, but they are effective as initial guidelines for interpreting paleoclimate. For example, Pratt (1984) found that discrete illite is much more abundant in the dark calcareous shales of the Hartland Shale Member of the Greenhorn Formation than in the limestone beds of the Bridge Creek Limestone Member. She attributed the stratigraphic changes in illite distribution to greater river discharge during deposition of the Hartland Shale. Leckie et al. (1991) documented a large increase in the proportion of kaolinite in the Lohali Point section during early Turonian time that coincides with peak transgression of the Greenhorn Sea. They suggested that regional climate became warmer and wetter as sea level rose, facilitating the formation of kaolinite adjacent to the seaway. Sethi and Leithold (1994) found cyclic changes in the relative abundance of clay minerals in the lower Turonian Mancos Shale and Tropic Shale in Utah. They noted an increase of mixed-layer illite-smectite in the more carbonate-rich beds that presumably formed during drier climatic periods.

In a core and in outcrop sections of the Greenhorn Formation near Pueblo, Colorado, a heavy trend in whole-rock $\delta^{13}C_{carb}$ values closely tracks the signal for organic matter ($\delta^{13}C_{org}$) through the Cenomanian-Turonian boundary interval of the lower Bridge Creek Member (Pratt et al., 1993) (Fig. 2). The limestone beds record the most negative $\delta^{13}C_{carb}$ values in the cyclic lithofacies of the Bridge Creek. Barron et al. (1985) proposed that the lighter $\delta^{13}C_{carb}$ values of the limestones reflect times of greater mixing and carbonate productivity due to availability of isotopically light and nutrient-rich deep waters. Eicher and Diner (1989) also support an interpretation of increased productivity during limestone deposition based on the presence of abundant calcispheres, which indicate high productivity, in the limestone beds; the concentration of clay-size quartz in many of the limestones, which are believed to be derived from the dissolution of radiolarian skeletons, and contrasts in $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ between the limestones and adjacent marlstones.

Eicher and Diner (1985, 1989) propose that the rhythmic bedding of the Bridge Creek Member is the result of productivity cycles rather than dilution cycles, which were driven by periodic changes in water mass production and vertical mixing. These authors emphasize that salinity-sensitive planktic foraminifera characterize the marlstones as well as the limestones. They reject the hypothesis that the marlstones represent times of increased runoff with the attendant development of a sediment-laden brackish cap across the seaway. In support of this argument, molluscan trophic and species diversity - including ammonites, epifaunal and infaunal bivalves, and gastropods - are greatest along the western clay-dominated corridor of the seaway (northeastern Arizona and southern Utah) where the brackish cap would be best developed, compared with the transitional and carbonate-dominated central sectors (Elder, 1990, 1991; Kirkland, 1991).

These observations suggest that if precipitation and runoff from the Sevier orogenic belt or any other area increased during the wetter part of the climate cycle, when the marlstone or calcareous shale were being deposited, then only minor salinity changes in the upper water column are feasible given the stenohaline nature of most marine organisms. Pratt et al. (1993) suggested that the amplitude of $\delta^{18}O_{carb}$ change ($2\text{-}3^{o}/_{oo}$) across the limestone-shale couplets translates into salinity variability of 10-20% (i.e., $3.4\text{-}7.0^{o}/_{oo}$). Fischer et al. (1985) concluded that the rhythmic bedding of the Bridge Creek Member was due to a combination of dilution and carbonate productivity cycles. We concur with this interpretation and suggest that the darker marlier and shalier parts of the bedding couplets - those lithofacies examined in this study - accumulated under minor reductions in near-surface water salinities ($<7^{o}/_{oo}$). Small salinity gradients between the surface and bottom in continental shelf or semi-enclosed sea settings are sufficient to stratify a relatively stable water column and therefore influence benthic ventilation (Mann and Lazier, 1991; Jewell, 1993).

Oceanography

A global positive excursion in whole-rock carbonate $\delta^{13}C$ values is well-developed in uppermost Cenomanian-basal Turonian marine strata of the U.S. western interior (Pratt and Threlkeld, 1984; Barron et al., 1985; Pratt, 1985; Arthur et al., 1987, 1988, 1990; Schlanger et al., 1987; Pratt et al., 1993). This excursion lasted about 0.5 m.y. and is attributed to enhanced marine productivity and the burial and preservation of organic matter with rising global sea level. These conditions correspond to times of increased low-latitude deep water production on flooded continental margins and heightened intraplate and convergent margin volcanism. Sediments of the North Atlantic also record dramatic changes in deep water ventilation and carbonate preservation related to the opening of the deep water gateway between the North and South Atlantic at this time (Tucholke and Vogt, 1979; Leckie, 1989). Heightened productivity expanded oxygen minimum zones (OMZ) in the mid- to upper water column along continental margins and under areas of oceanic divergence. Oxygen-poor water masses may have invaded epicontinental seas with rising sea level, enhancing preservation of organic matter (Arthur et al., 1987; Schlanger et al., 1987; Slingerland et al., 1996). Frush and Eicher (1975) proposed that the incursion of an OMZ into the Greenhorn Sea may have limited benthic foraminifera through much of the upper Cenomanian-lower Turonian strata of the deep central axis of the seaway. An alternate hypothesis is that the mixing of polar waters and subtropical waters in the Western Interior Sea produced its own oxygen-depleted, intermediate water mass through cabbeling (Hay et al., 1993; Fisher et al., 1994).

Parrish et al. (1984) suggested that oceanographic circulation in the southern part of the Western Interior Sea may have alternated seasonally between north-flowing and south-flowing, whereas Jewell (1993) proposed that a wind-driven subtropical gyre dominated flow south of latitude 50^{o}N; the latter implies northward flow along the western side of the seaway and southward flow along the eastern side. In contrast, the distribution of planktic foraminifera in the Cenomanian-Turonian Greenhorn Sea suggests that warm, normal marine waters moved northward along the central and eastern sides of the seaway during transgression and cooler, slightly less saline waters moved southward along the western side (Eicher, 1969a; Eicher and Diner, 1985; Fisher et al., 1994). Numerical simulations of wind-driven circulation in the seaway suggest that oceanic circulation was dominated largely by storms, particularly winter storms, and that storm-driven shelf currents on the western side of the seaway were predominantly to the south (Ericksen and Slingerland, 1990).

Circulation experiments using a global climate model (GENESIS) and a coastal ocean circulation model (CIRC) indicate that circulation in the early Turonian Greenhorn Sea was driven largely by freshwater runoff (Arthur et al., 1996; Slingerland et al., 1996). Runoff from the eastern side of the seaway flowed northward as a coastal jet and runoff from the western side flowed southward. The seaway exported freshwater and drew in both Boreal and Tethyan waters, creating a strong counterclockwise gyre (Slingerland et al., 1996). These model results are in good agreement with planktic and benthic foraminiferal distribution patterns (Eicher and Diner, 1985; Caldwell et al., 1993; Fisher et al., 1994; West et al., this volume).

PLANKTIC FORAMINIFERAL PALEOECOLOGY AND PALEOBIOGEOGRAPHY IN THE WESTERN INTERIOR SEA

Sediment assemblages of fossil planktic foraminifera provide useful information about the nature of the ancient uppermost water column, such as its stratification and productivity. Most species of modern planktic foraminifera are adapted to relatively narrow ranges of temperature and salinity (Be, 1977); this relationship that holds true for most marine organisms and probably for Cretaceous planktic foraminifera as well. The richness of species, genera, or morphotypes is greatest in well preserved sediment assemblages deposited under normal salinity, generally low-nutrient waters of the low- to mid-latitudes where seasonal or year-round temperature gradients in the upper water column provide a variety of trophic and density-specific niches (Lipps, 1979; Leckie, 1989; Hallock et al., 1991) (Fig. 3). Sediment assemblages that accumulate under ecotones - areas of the ocean where two surface water masses meet and co-mingle - may actually have slightly higher simple diversity due to mixing of biocenoeses (Cifelli and Benier, 1976; Hallock et al., 1991). Stratification in the upper water column, which is caused by seasonal changes in the strength or position of the thermocline, is known to be fundamentally important in maintaining marine plankton communities. It affects, for example, nutrient availability and recycling, productivity, seasonal succession, reproduction, and predation (Mann and Lazier, 1991).

The relative abundance of planktic foraminifera to total foraminifera in depositional settings above the lysocline and carbonate compensation depth are very useful in delineating water depth or surface water productivity. Planktic foraminifera are rare in nearshore environments but increase in abundance relative to benthic taxa across the continental shelf (e.g., Phleger, 1951; Bandy, 1956; Murray, 1976; Gibson, 1989). In the modern ocean, planktics typically constitute 90-99% of outer neritic to mid-bathyal foraminiferal sediment assemblages (Fig. 3). However, elevated surface water productivity can significantly reduce the relative abundance of planktics. The enhanced flux of organic matter from the surface waters to the seafloor stimulates benthic productivity, and while the flux of planktic foraminiferal shells too may be higher, the greatest increase in relative abundances is in the benthic foraminifera and other benthic organisms, for example, ostracods, echinoderms, and sponges (Leckie, 1987; Berger and Diester-Haass, 1988; Herguera and Berger, 1991).

During the mid-Cretaceous, biserial (*Heterohelix*) and triserial (*Guembelitria*) planktic morphotypes were much more common in the shallower, more proximal waters of epicontinental seas than in the more distal, open ocean settings that were their habitat in the later Cretaceous and Cenozoic (Leckie, 1985; 1987; Nederbragt, 1990). For example, Tappan (1940) and Mancini (1982) noted the common occurrence of *Guembelitria* and *Heterohelix* in the warm, neritic environment of the lower Cenomanian Grayson Formation of north-central Texas. *Heterohelix* was widely distributed and was one of the most abundant planktic foraminiferal groups in the seaway during Cenomanian-Turonian time (e.g., Eicher, 1969a; Caldwell et al., 1978; McNeil and Caldwell, 1981), and was reported as far north as the Arctic Slope of Alaska (Tappan, 1962). On the other hand, *Guembelitria* was more restricted to southern and presumably warmer waters, and its greatest relative abundances are reported from neritic localities (e.g., Hazenbush, 1973; Lessard, 1973; Mancini, 1982; Leckie et al., 1991; Olesen, 1991). On the western side of the seaway, *Guembelitria* has been reported from as far north as southeastern Utah, where it is common to abundant at some levels (Lessard, 1973), and in the central part of the seaway, *Guembelitria* has been found in calcareous shales as far north as the Black Hills (Fisher, pers. commun., 1995). Forms referable to *Guembelitria* sp. from the prairie provinces of Canada (e.g., Stelck and Wall, 1954, 1955; Wall, 1960, 1967; Caldwell et al., 1978) may prove to be a benthic buliminid.

Leckie (1987) called *Heterohelix* and *Guembelitria*, among others, the "Epicontinental Sea Fauna" on the basis of their characteristic mid-Cretaceous distribution pattern. Perhaps these genera, particularly the earliest species, had a benthic stage to their life-cycle thereby placing water depth constraints on their distributions. Alternatively, perhaps these genera were broadly tolerant (eurytopic) of variable or unstable neritic waters with ex-

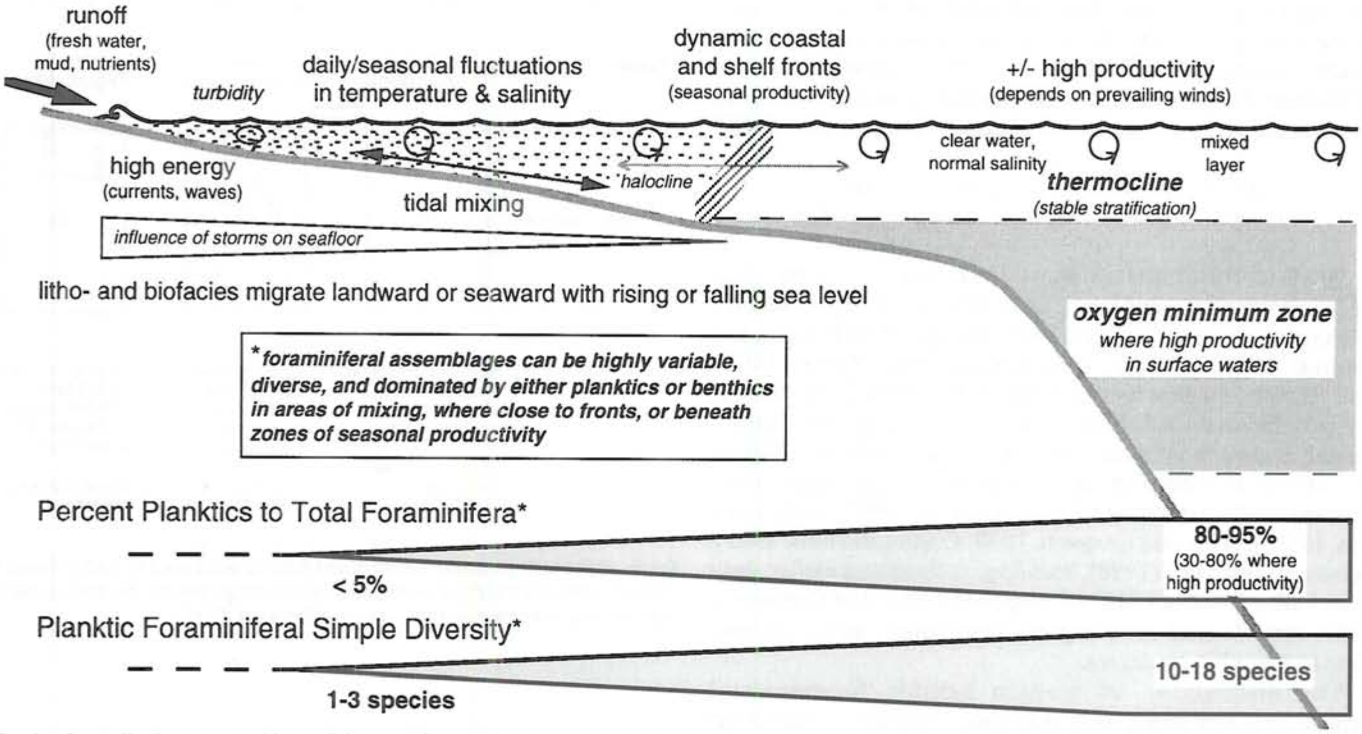

FIG. 3—Generalized oceanographic conditions and foraminiferal assemblage characteristics in neritic to upper bathyal settings along a dominantly siliciclastic margin. Offshore trends in the proportion of planktic to total foraminifera and simple planktic diversity apply to the Cenomanian-Turonian (see text for details).

tremes of seasonal productivity, salinity, and/or temperature (Fig. 3).

The distribution of Cretaceous and Cenozoic biserial and triserial planktic foraminifera has been enigmatic compared with most trochospirally-coiled species. For example, *Heterohelix* proliferated with the expansion of epicontinental seas during the late Cenomanian and early Turonian, and during the time of OAE-2 (e.g., Eicher, 1969a; Sliter, 1972; Jarvis et al., 1988; Nederbragt, 1990). Its Cenozoic biserial look-alikes, *Chiloguembelina* and *Streptichilus* may have been associated with low oxygen waters in the lower part of the seasonal thermocline (Boersma and Premoli-Silva, 1989; Resig, 1993). *Gallitellia*, the present-day homeomorph of the triserial *Guembelitria*, is very sparse in the modern ocean but occurs in fairly significant abundances in the Red Sea and in upwelling areas of the northern Indian Ocean (Kroon and Nederbragt, 1990).

Trochospiral morphotypes, particularly species of *Hedbergella* and *Whiteinella*, were the "weeds" of the mid-Cretaceous oceans, found in great abundances in both epicontinental sea and open ocean settings. These taxa are grouped as the major components of the "Shallow Water Fauna" because, like the more common and widespread species of the oceans today, they probably lived in the sun-lit near-surface waters of the mixed layer and/or upper reaches of the thermocline, where trophic resources are concentrated (Leckie, 1987, 1989). Along with the biserial forms, the trochospiral morphotypes were among the most abundant and widespread planktics in the Greenhorn Sea, especially *Hedbergella delrioensis*, *H. planispira*, and *H. loetterlei* (Eicher, 1969a; Eicher and Worstell, 1970; Caldwell et al., 1978; McNeil and Caldwell, 1981; Schroeder-Adams et al., 1996). *Hedbergella loetterlei* is also reported from the north slope of Alaska (Tappan, 1962). *Globigerinelloides*, a planispiral morphotype, occurs widely, although they are rarely common and tend to disappear before *Hedbergella* or *Heterohelix* in the shoreward direction. Other mid-Cretaceous genera such as *Clavihedbergella* and *Schackoina* have distributions and relative abundances similar to *Globigerinelloides*. Species of all these genera ranged as far north as Alberta, Saskatchewan, and Manitoba at the time of peak transgression of the Greenhorn Sea during the early Turonian (e.g., Caldwell et al., 1978; McNeil and Caldwell, 1981). These last three taxa are also included in the "Shallow Water Fauna" based on their paleobiogeography, but they were probably more stenotopic than some species of *Hedbergella* and *Whiteinella*.

Cenomanian-Turonian keeled morphotypes, species of *Rotalipora*, *Praeglobotruncana*, *Dicarinella*, and *Marginotruncana*, are typically diagnostic components of high-diversity assemblages and indicate warm, normal marine, stratified water masses (Leckie, 1987, 1989). These taxa are collectively grouped as the "Deep Water Fauna", not necessarily because they were the deepest-dwelling as originally proposed by Leckie (1987), but because they were more common in deeper, more distal 'blue water" settings, and because they were probably the most stenotopic planktic foraminifera of this age. As a group, keeled species probably occupied a diverse range of habitats from the mixed layer to various depths along the thermocline, or perhaps even deeper (e.g., Douglas and Savin, 1978; Hart, 1980; Caron and Homewood, 1982; Jarvis et al., 1988; Corfield et al., 1990). In the Greenhorn Sea, this group of taxa had the most limited biogeographic range of all planktic foraminifera (e.g., Eicher, 1969a). Keeled taxa are not found north of the Black Hills (Fisher, pers. commun., 1996). Their distribution was restricted primarily to the eastern half of the seaway, centered on the axial part of the basin in eastern Colorado (Eicher, 1969a; Eicher and Diner, 1985, 1989; Fisher et al., 1994). Eicher (1969a) reported *Rotalipora* from west-central Colorado and Hazenbush (1973) reported both *Rotalipora* and *Praeglobotruncana* from northeastern Arizona, but these and other keeled taxa are only represented by two specimens in the Cenomanian-Turonian boundary interval at our MV section in the southwestern corner of Colorado. No keeled taxa were reported from the Red Wash section in the northwestern corner of New Mexico (Lamb, 1968).

BENTHIC FORAMINIFERAL PALEOECOLOGY AND PALEOBIOGEOGRAPHY IN THE WESTERN INTERIOR SEA

Benthic foraminifera have long been used to infer paleoenvironmental conditions, including water depth, salinity, and oxygenation (e.g., Phleger, 1951; Bandy, 1956; Douglas and Woodruff, 1981; Bernhard, 1986; Culver, 1988; Murray, 1991). More recently, modern benthic foraminifera have been found to occupy different microhabitats - for example, epifaunal, shallow infaunal, and deep infaunal - and there is a significant relationship between microhabitats and foraminiferal test shape, mode of coiling, and distribution of pores (Corliss, 1985; Corliss and Chen, 1988; Corliss and Emerson, 1990; Corliss and Fois, 1991). Koutsoukos and Hart (1990), building on these and earlier studies, including Chamney (1976), proposed a model of Cretaceous benthic foraminiferal morphogroup distribution, paleo communities, and trophic strategies.

The distribution of modern benthic foraminiferal morphotypes, and hence microhabitats, depends on numerous factors including food supply and oxygen content below the sediment/water interface (Corliss and Emerson, 1990; Jorissen et al., 1995) (Fig. 4). For example, Gooday (1993) and Thomas and Gooday (1996) have suggested that some epifaunal taxa are trophic opportunists that respond quickly to the input of food, such as seasonal pulses of phytodetritus from the surface waters associated with the spring bloom. In addition, there is growing evidence that some infaunal foraminiferal taxa may have a competitive advantage in dysoxic or anoxic environments by requiring very little oxygen (microaerophiles) or being capable of surviving without oxygen for an extended interval of time (facultative anaerobes), or by harboring symbiotic chemoautotrophic bacteria (West, 1993; Bernhard, 1996).

The Cretaceous of the western interior contains a rich benthic foraminiferal record. Cenomanian-Turonian foraminifera of the United States and Canada have been well documented (Tappan, 1940; Cushman, 1946; Young, 1951; Jones, 1953; Fox, 1954; Frizzell, 1954; Stelck and Wall, 1954, 1955; Wall, 1960, 1967; Eicher, 1965, 1966, 1967; Lamb, 1968; Eicher and Worstell, 1970; Hazenbush, 1973; Lessard, 1973; Frush and Eicher, 1975; North and Caldwell, 1975; McNeil and Caldwell, 1981; Eicher and Diner, 1985; Leckie, 1985; Bloch et al., 1993; Caldwell et al., 1993; Fisher et al., 1994; and Schroeder-Adams et al., 1996). Aspects of their spatial distribution have been summarized by Caldwell et al. (1978, 1993) and Eicher and Diner (1985). Benthic foraminiferal faunas from the Canadian prairie provinces and from sections in Texas differ significantly from each other, not only in terms of species present, but also in gross population structure and diversity. For example, northern Cenomanian-Turonian assemblages are typically dominated by agglutinated taxa while coeval southern assemblages are dominated by calcareous taxa, both numerically and in species richness. Therefore, the paleobiogeography of calcareous and agglutinated benthic taxa offer an important, yet largely underutilized, proxy for water mass affinities in the Cretaceous Western Interior Sea.

Eicher and Worstell (1970) documented the taxonomy and distribution of foraminifera from eight Cenomanian-Turonian localities of the Great Plains (Table 1; Fig. 1); four northern sections from around the Black Hills of northeastern Wyoming and southwestern South Dakota (localities 1-4) and four southern sections from eastern Colorado and western Kansas (localities 5-8). All sections lie near the axis of the seaway and are composed of calcareous shales and/or marlstones in the Cenomanian-Turonian boundary interval; interbedded limestones occur in all sections but the northernmost in Bull Creek Wyoming. This section is very close to the western edge of a pronounced regional facies change from calcareous to noncalcareous shales (Eicher and Worstell, 1970; Fisher et al., 1994).

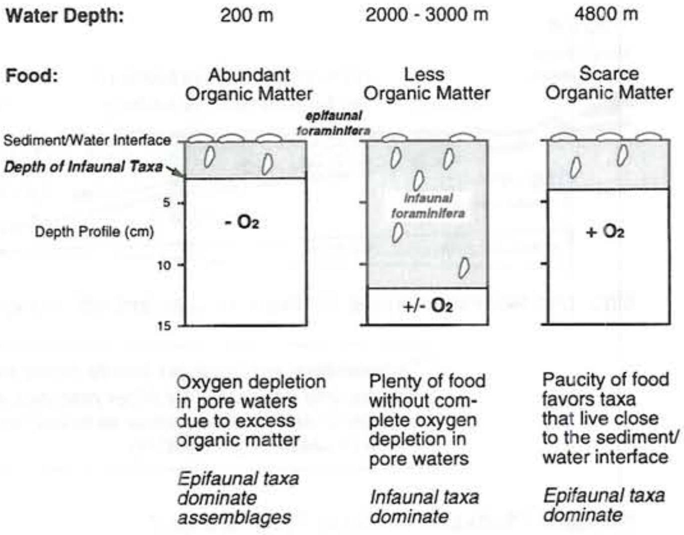

FIG. 4—Relationship between epifaunal and infaunal dominance in benthic foraminiferal assemblages and the amount of organic matter supplied to the seafloor (summarized from data presented by Corliss and Emerson, 1990).

Eicher and Worstell (1970) recognized three major assemblage zones in these strata: (1) the upper Cenomanian "lower planktonic zone", in which planktics dominate to the near-exclusion of benthics, (2) the upper Cenomanian "benthonic zone", which is characterized by abundant planktic foraminifera and diverse populations of benthic foraminifera dominated by calcareous taxa, and (3) the uppermost Cenomanian-middle Turonian "upper planktonic zone", which is characterized by high dominance, very low diversity planktic assemblages, and often sparse benthic assemblages. The lower boundary of the benthonic zone is sharp at the southern sections, but it is gradual in the northern sections. The paucity of benthic foraminifera in strata from below and above the "benthonic zone" was attributed to oxygen-poor conditions on the seafloor (Eicher and Worstell, 1970; Frush and Eicher, 1975; Eicher and Diner, 1985).

The benthonic zone represents an important oceanographic event within the Greenhorn Sea. The biggest change in benthonic zone assemblages occurs between the southern Black Hills and east-central Colorado (Fig. 1). Most of the agglutinated species documented from the benthonic zone have northern affinities based on their distributions across the northern Great Plains of the United States and Alberta (Fox, 1954; Stelck and Wall, 1955; Eicher, 1967); most notable of these are *Ammobaculoides mosbyensis*, *Trochammina wetteri*, *Reophax recta*, and *Haplophragmoides topagorukensis* (Table 1). *Spiroplectinata vulpes* occurs as far south as Rock Canyon but has greater relative abundances in the northern sections, which also suggests northern affinities. *Textularia rioensis* and *Gaudryina* cf. *G. quadrans*, on the other hand, are absent from the northern Black Hills sections but are characteristic, albeit very rare, of the benthonic zone of eastern Colorado and western Kansas (Eicher and Worstell, 1970). Southern affinities for *T. rioensis* are strongly supported by its common occurrence in the Grayson Formation of north-central Texas (e.g., Tappan, 1940; Mancini, 1982).

Numerous calcareous benthic taxa are widely distributed across the Great Plains (Eicher and Worstell, 1970), including: *Quinqueloculina moremani*, *Citharina kochii*, *Dentalina basiplanata*, *D. communis*, *Lenticulina gaultina*, *Ramulina aculeata*, *Buliminella fabilis*, *Neobulimina albertensis*, *Tappanina laciniosa*, *Valvulineria loetterlei*, *Pleurostomella nitida*, *Fursenkoina croneisi*, *Cassidella tegulata*, *Gavelinella dakotensis*, *G. plummerae*, *Lingulogavelinella asterigerinoides*, *L. modesta*, *L. newtoni*, and *Orithostella viriola* (Table 1). Of these, *Tappanina laciniosa* and *Lingulogavelinella newtoni* are more abundant in the southern sections, and *Valvulineria loetterlei* and *Pleurostomella nitida* are more common in the Kansas sections (Table 1). We propose that these last four taxa have southern, Gulf Coast affinities and are particularly diagnostic of "healthy" benthic conditions: abundant food, ample supply of dissolved oxygen, and warm, clear, and well-circulated waters.

Buliminella fabilis and *Gavelinella plummerae* are most abundant in eastern Colorado and westernmost Kansas. We suggest that these two species are diagnostic of the deepest part of the seaway and are either intermediate water mass or bathymetric indicators (Table 1). To further support the interpretation of Gulf Coast affinities for all of these taxa, we also note that *Valvulineria loetterlei*, *Gavelinella plummerae*, and *Lingulogavelinella asterigerinoides* are common to abundant constituents of the lower Cenomanian Grayson Formation of north-central Texas (Tappan, 1940; Mancini, 1982). In addition, these three taxa plus *Buliminella fabilis*, *Lingulogavelinella modesta*, *Lenticulina gaultina*, and *Neobulimina albertensis* are among nine species of calcareous benthics associated with a brief incursion of warm southern waters into southern Colorado (Thatcher Limestone Member of Graneros Shale) during early Cenomanian time (Eicher, 1965; Eicher and Diner, 1985).

RESULTS

Planktic Foraminifera

The relative abundance of planktic foraminifera (percent planktics to total foraminifera; Fig. 5) increases in the offshore direction from LP (neritic) to MV (proximal basin) to RC (distal basin). The section at RC is dominated by planktics throughout the study interval with an average of 89.5%, a typical value for outer neritic-upper bathyal settings. The sections at LP and MV show a marked decline in planktic foraminifera in the uppermost Cenomanian *Neocardioceras* zone. In all three sections there is a small to moderate drop in planktics for a brief interval within the upper Cenomanian *Sciponoceras* zone.

The section at LP is dominated by biserial (*Heterohelix*) and triserial (*Guembelitria*) taxa, except in the upper Cenomanian *Metoicoceras* zone, where trochospiral taxa (mostly *Hedbergella*) dominate and there is an influx in planispiral forms (*Globigerinelloides*) (Fig. 6). Although we did not find any specimens of keeled species in our samples, Hazenbush (1973) reported *Rotalipora greenhornensis* and *Praeglobotruncana stephani* in two samples from the upper Cenomanian of LP. RC is dominated by trochospiral and biserial species. Significant numbers of planispiral forms are present in the *Sciponoceras* zone. Triserial forms are very rare, but seven keeled species, including *Rotalipora cushmani* and *R. greenhornensis*, are present in the upper Cenomanian of the RC section (Leckie, 1985).

The sections at both LP and RC show an abrupt increase in biserial forms at the base of the uppermost Cenomanian *Neocardioceras* zone. This dramatic change in planktic foraminiferal assemblages ("*Heterohelix* shift") roughly coincides with a major $\delta^{18}O$ depletion event (Pratt, 1985). At Blue Point, a more proximal section in the Black Mesa basin, Olesen (1991) also documented a marked increase in *Heterohelix* in the uppermost Cenomanian. The planktic foraminiferal assemblages at MV are transitional in overall composition between LP and RC, particularly in their relative abundance of the triserial *Guembelitria*. Importantly, however, population structure through the MV section is highly variable from sample to sample, the *Heterohelix* shift is weakly developed, and only two specimens of keeled taxa were found in the upper Cenomanian strata (West et al., 1990).

Benthic Foraminifera

Benthic foraminiferal assemblages at LP and MV are very similar. There is a significant proportion of agglutinated taxa with northern affinities (Fig. 7). Both sections also display nearly identical stratigraphic trends in benthic assemblage changes. By contrast, the upper Cenomanian *Sciponoceras* zone at RC is dominated by a diverse assemblage of calcareous benthics (the "benthonic zone" of Eicher and Worstell (1970)), many of which have southern affinities. A majority of these calcareous species are not present at either LP or MV. In addition, the RC section only contains trace specimens of two agglutinated species, and these have southern affinities.

All three sections show a marked increase in the relative abundance of *Gavelinella dakotensis*, a presumed epifaunal taxon, in the uppermost Cenomanian *Neocardioceras* zone. This is followed by a rapid shift to dominance by *Neobulimina albertensis*, a presumed infaunal taxon, within or at the top of the *Neocardioceras* zone (Cenomanian-Turonian boundary). Note that this abrupt change in benthic assemblages appears to be diachronous from basin center to western margin, occurring first at RC within the *Neocardioceras* zone and last at LP at the Cenomanian-Turonian stage boundary.

In summary, the planktic assemblages at LP are similar to those at RC, despite being some 450 km apart and representing neritic and deep water environments, respectively, but the LP benthics are more similar to those at MV. Additional environmental constraints, such as clay mineralogy, are needed to interpret this pattern.

Clay Mineralogy

The clay mineralogy of the sections at LP and RC are very similar and display identical stratigraphic trends through the Cenomanian-Turonian boundary interval (Fig. 8). In unadjusted diffractograms, mixed-layer illite/smectite is the prominent clay at both sites, with lesser amounts of discrete illite and kaolinite. By contrast, illite more obviously contributes to the MV clay assemblages, but kaolinite and illite/smectite are major components

TABLE 1—BENTHIC FORAMINIFERA OF THE BENTHONIC ZONE[1]

Localities	**1**	**2**	**3**	**4**	**6**	**RC**	**7**	**8**	**LP**	**MV**
	Bull Creek	Belle Fourche	Black Gap	Hot Springs	Rock Canyon	Rock Canyon	Hartland-BC	Bunker Hill	Lohali Point	Mesa Verde
	NE Wyoming	W S.Dakota	W S.Dakota	SW S.Dakota	S Central Colo.	S Central Colo.	SW Kansas	N Central Kan.	NE Arizona	SW Colorado
Agglutinated Species										
Saccammina alexanderi			0.1						0.2	
Reophax recta	2.3								0.7	
Haplophragmoides topagorukensis	0.0	0.1	0.0						0.6	1.4
Haplophragmium sp.				1.0	0.1					
Ammobaculites junceus	0.0			0.2						
Ammobaculoides mosbyensis	11.6								0.2	
Coscinophragma? codyensis	0.2	0.2	0.1	1.5					1.3	
Textularia rioensis			0.0		0.2	0.1	0.6			
Trochammina rainwateri	0.3		0.0						5.6	0.9
Trochammina wetteri	2.5	0.1	1.1						8.6	3.7
Gaudryina cf. G. quadrans					0.2		0.6	0.1		
Verneuilina sp.	0.0									
Spiroplectinata vulpes	1.7		1.2	0.3	0.1					
Marssonella conica	0.0		0.0							
Ammobaculites impexus									0.4	0.4
Ammobaculites fragmentarius									0.2	
Trochamminoides apricarius									1.9	2.1
Haplophragmoides spp.									0.6	
Calcareous Species										
Quinqueloculina moremani	0.3		0.2		0.1	0.1	0.4		2.2	3.4
Massilina planoconvexa			0.0							
Citharina complanata			0.0		0.1		0.4	0.1		
Citharina kochii	0.3	0.5	0.2	0.7	0.2	0.3	0.6	0.4	1.3	0.1
Citharina petila	0.1		0.1		0.1					
Dentalina basiplanata	0.1			0.3	0.1				0.7	0.1
Dentalina communis	0.2	0.5	0.2	0.2	0.2	0.1	0.2			
Dentalina intrasegma	0.0		0.1		0.2		0.4	0.1		
Frondicularia extensa			0.0				0.2			
Frondicularia imbricata	0.1	0.1	0.0	0.2	0.1					
Lagena apiculata	0.1									
Lagena striatifera	0.0		0.0							
Lagena sulcata	0.1									
Lenticulina gaultina	0.4	0.5	0.1	0.2	0.3	0.3	0.8	0.5	2.1	
Marginulina siliquina	0.0									
Marginulinopsis amplaspira	0.0	0.2				0.1				
Nodosaria bighornensis	0.3		0.2	0.2			0.2		0.9	
Planularia dissona	0.0									
Pseudonodosaria sp.	0.1		0.1				0.2			
Saracenaria reesidei	0.1							0.1		
Vaginulina debilis	0.1		0.0				0.2			
Vaginulina cretacea	0.1									0.1
Lingulina nodosaria	0.1		0.0		0.1		0.4			
Globulina lacrima	0.1		0.0		0.2			0.1		
Pyrulina cylindroides	0.2		0.0			0.1	0.2			
Bullopora laevis	0.1						0.2			
Ramulina aculeata	0.1		0.1		0.1	0.1		0.1		
Ramulina globulifera	0.1		0.0				0.2			
Washitella sp.	0.0									
Buliminella fabilis	14.8	7.7	11.4	12.2	25.7	30.3	20.8	9.6		0.3
Neobulimina albertensis	28.7	50.3	48.0	36.7	18.4	14.3	7.8	4.5	57.5	57.4
Tappanina laciniosa	6.9	6.0	15.3	13.6	19.3	20.3	31.9	56.5		
Valvulineria loetterlei	5.9	0.3	5.0	2.2	6.8	4.6	11.1	11.5		
Spirillina minima	0.1			0.2						
Pleurostomella nitida	0.8		1.2	0.2	0.4	1.8	2.1	3.8		
Fursenkoina croneisi	0.0		0.6	0.2	0.4	0.2	0.2	0.1		
Cassidella tegulata	0.3	0.2	0.6	1.0	0.4	0.2	0.6	0.1		
Gavelinella dakotensis	10.5	32.8	5.1	21.6	6.1	3.4	0.2	0.2	15.0	22.1
Gavelinella plummerae	2.4		0.2	1.2	12.0	13.5	7.3	1.1		
Lingulogavelinella asterigerinoides	1.9	0.2	0.3	0.9	1.5	5.5	1.0	2.8		
Lingulogavelinella modesta	1.0		7.0	3.6		0.7	0.6	0.5		
Lingulogavelinella newtoni	0.2	0.1	0.9	1.5	6.5	4.0	10.7	7.2		
Orithostella viriola	0.7		0.0	0.2	0.1		0.2			
Conorboides minutissima	0.3		0.2			0.2		0.4		
Hoeglundina charlottae	3.1					0.1				0.1
Number of Samples in Analysis	15	6	8	4	4	7	3	3	6	6
Total Number of Specimens	4129	860	2333	588	1139	1989	523	819	534	700

[1] Average benthic foraminiferal percentage data from the upper Cenomanian "benthonic zone" of Eicher and Worstell (1970) from across the U.S. western interior to illustrate paleobiogeographic affinities and water depth. Data are from Eicher and Worstell (1970; localities 1-8), Leckie (1985) for the RC section, and this study. Refer to Fig. 1 for section locations.

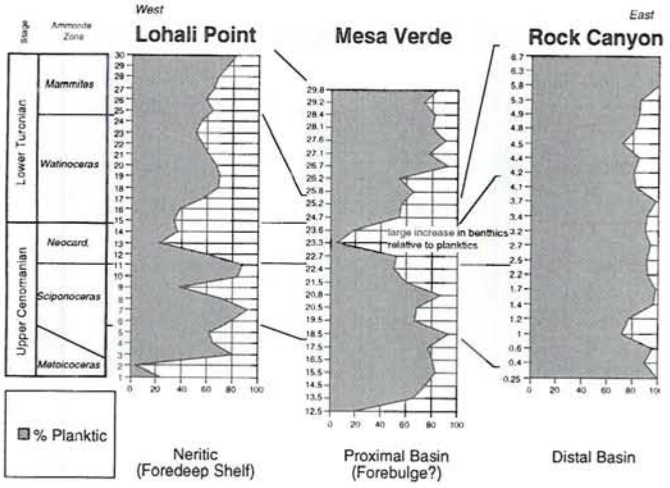

FIG. 5—Proportion of planktic taxa to total foraminifera through the Cenomanian-Turonian boundary interval of Lohali Point (LP), Mesa Verde (MV), and Rock Canyon (RC). Sample numbers for the LP and MV sections of the Mancos Shale correspond to meters above the Dakota Sandstone. Sample numbers for the RC section of the Greenhorn Formation correspond to meters above the base of the Bridge Creek Limestone Member (-0.25 m is from the uppermost Hartland Shale Member). Note the major drop in planktics and increase in benthics in the *Neocardioceras* zone (uppermost Cenomanian) of the LP and MV sections.

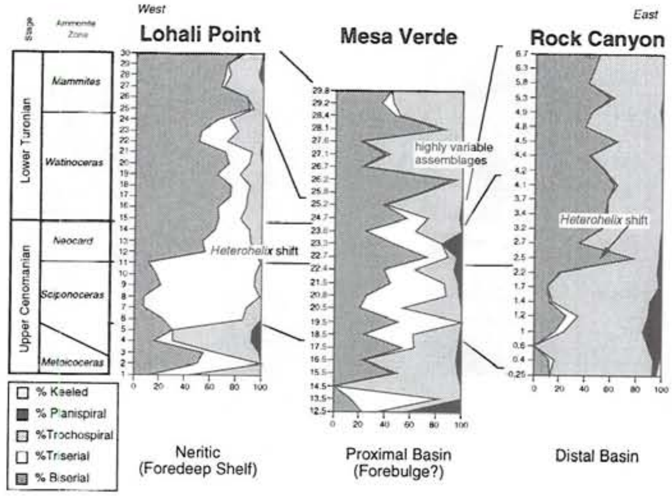

FIG. 6—Major planktic foraminiferal morphotypes. Note large sample-to-sample variability at the MV section compared with LP section to the west and RC section to the east. Also note the abrupt change in planktic populations (*Heterohelix* shift) at the base of the *Neocardioceras* zone at both LP and RC. Refer to Fig. 5 caption for additional explanation.

as well. Of particular interest is the chlorite in the section at MV; this clay mineral is not found at the other sites. Chlorite is found throughout the section with a relatively stable peak intensity. Although these intensity values do not represent absolute mineral abundance, the trends depicted are reproduced when the data are adjusted according to the mineral intensity factors (MIF) in the manner proposed by Reynolds (1989).

The stratigraphic fluctuations in relative clay-mineral abundance are most pronounced in the LP section, with the abundance of mixed-layer illite/smectite increasing from the base of the section up through the middle of the *Sciponoceras* zone. This trend reflects westward shoreline migration with transgression of the Greenhorn Sea. A similar, albeit more subdued expression of the same trend is visible in the RC data. At both sites, the *Sciponoceras-Neocardioceras* zone transition is marked by a doublet peak of increased kaolinite and illite. The upper peak, near the base of the *Neocardioceras* zone, is more intense, particularly in the deeper-water, distal RC section. Kaolinite and illite decline at the Cenomanian-Turonian boundary, then increase again in the lower Turonian. This latter effect is more pronounced at the more proximal LP section. The clay-mineral abundance pattern at MV does not show the same degree of synchroneity, suggesting that major depositional processes at this site differed from those at the other two sites.

DISCUSSION

Foraminiferal Assemblages

The high, relatively stable abundances of planktics through the Cenomanian-Turonian boundary interval at the RC section, coupled with its high planktic species richness, especially in the *Sciponoceras* zone, indicate a fully marine, deep water, stratified environment (Fig. 5). The section at RC contains distal epicontinental sea assemblages of planktic foraminifera, with an abundance and diversity of trochospiral, biserial, keeled, and planispiral taxa, and by paucity of triserial forms. The central core of the seaway was essentially an arm of the Tethys Sea (Eicher and Diner, 1989). The southwestern side of the Greenhorn Sea contains warm water neritic assemblages based in part on high abundances of triserial and biserial taxa and paucity of keeled forms at LP (Fig. 6). The section at LP is also characterized by lower relative abundances of planktic foraminifera and thereby indicative of mid- to outer-neritic depths. Despite being located some 450 km apart and containing deep and shallow water assemblages of planktics, respectively, both RC and LP show a synchronous and marked change in population structure at the base of the uppermost Cenomanian *Neocardioceras* zone (Figs. 6, 9). This suggests that the *Heterohelix* shift represents a widespread oceanographic event in the southwestern Greenhorn Sea during latest Cenomanian time and it supports the hypothesis that the seaway at the RC and LP localities had the same surface water mass at this time (Leckie et al., 1991).

Although the section at MV contains abundant planktics and has compositional trends that are, in general, transitional between LP and RC, it also contains a number of important distinguishing features. First, the planktic assemblages vary greatly from sample to sample. Perhaps this is due to winnowing and current activity on a bathymetric high. If winnowing of foraminifera tests was responsible, then why don't the benthic populations or the ratio of planktics to benthics (% planktics) display as much variability, and why do the stratigraphic trends in benthics and in percent planktics parallel those at LP so well? Second, the *Heterohelix* shift is poorly developed at MV and very few warm water keeled species were recovered from the study interval (West et al., 1990). Again, perhaps this is a consequence of shallow water depths. If shallow depths were the cause, then why are the assemblages at MV, in general, transitional between LP and RC in both percent planktics and in overall population structure (Figs. 5, 6)? Water mass mixing or water column instability in a relatively shallow environment may have inhibited stable stratification or reduced surface water salinities (beyond tolerance of steno-

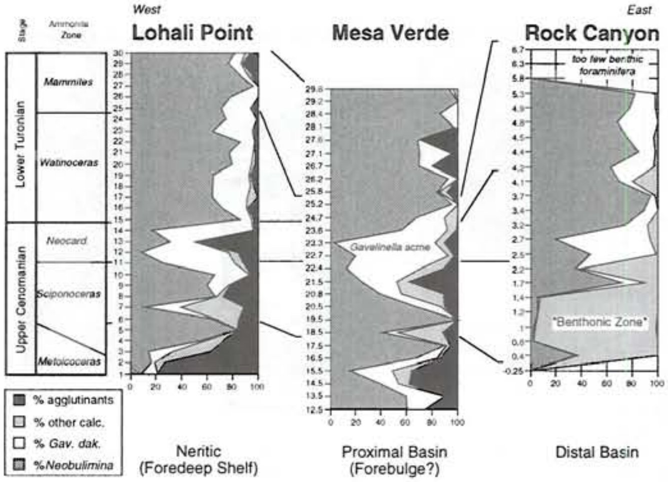

FIG. 7—Major benthic foraminifer morphogroups. Note striking similarity in the composition and stratigraphic trends of benthic foraminifers in the LP and MV sections, particularly in the upper Cenomanian. All three sites display an acme of *Gavelinella dakotensis* in the uppermost Cenomanian followed by an abrupt but diachronous change to *Neobulimina albertensis* dominance through the Cenomanian-Turonian boundary interval. Refer to Fig. 5 caption for additional explanation.

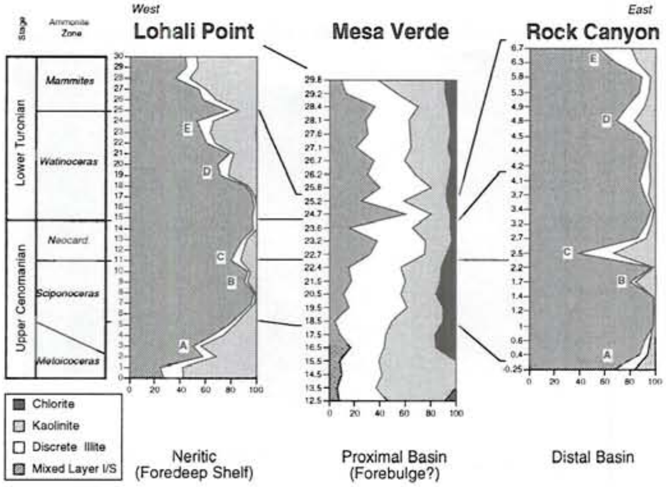

FIG. 8—Clay mineralogy. Letters A-E denote points of correlation between the sections at LP and RC. Note how very similar the composition and stratigraphic trends are between the LP and RC sections, while MV differs significantly from the other two. Refer to Fig. 5 caption for additional explanation.

haline taxa) thereby causing environmental exclusion of some taxa. Mixing of different water masses along a non-stationary oceanic front, or edge effect mixing and upwelling near MV, are two hypotheses that may explain both, the variable planktic assemblages and the weakly developed *Heterohelix* shift.

The benthos provide an entirely different paleoenvironmental view of change through the Cenomanian-Turonian boundary interval than the planktic record does. Based on the population structure of benthic foraminiferal sediment assemblages and their stratigraphic changes, it appears that LP and MV record very similar benthic environmental histories (Fig. 7). (1) Both sections display a marked increase in the proportion of benthics relative to planktics in the uppermost Cenomanian (Fig. 5), (2) both sections contain a distinct component of agglutinated taxa with northern affinities, which are completely lacking in the section at RC (Table 1), (3) both sections show nearly synchronous peak-to-peak changes in population structure through the upper Cenomanian, and (4) both sections lack the highly diverse calcareous assemblages of the benthonic zone (Eicher and Worstell, 1970) that are so well-developed and widespread in the deeper central and carbonate-dominated eastern parts of the seaway. Even the most diagnostic species of the benthonic zone (e.g., *Buliminella fabilis*, *Tappanina laciniosa*, *Valvulineria loetterlei*, *Gavelinella plummerae*, *Lingulogavelinella asterigerinoides*, *L. modesta*, and *L. newtoni*) are few or absent all-together at LP and MV (Table 1). We propose that the sections at LP and MV were situated at similar mid- to outer-neritic water depths and that both sites were exposed to a cool, northern water mass near the seafloor.

Clay Mineralogy

The presence of detectable chlorite in the MV section raises questions about the level of post-depositional thermal alteration of the clay minerals in that section (Fig. 8). Chlorite is commonly found as a product of burial diagenesis of sedimentary sequences. It usually appears during the conversion of mixed-layered illite-smectite to illite (Hower et al., 1976). However, several lines of evidence suggest a detrital origin for this chlorite. First, although ordered illite-smectite occurs in the lower 70 m of the 700 m Mancos Shale section at Mesa Verde (that is, in the upper Cenomanian-lower Turonian rocks that we studied) the bulk of the MV section is dominated by randomly interstratified illite-smectite, and chlorite persists through the entire section. This indicates that significant thermal diagenesis had not occurred in the MV study interval; illite and chlorite are not more abundant in the lower 70 m of the Mancos Shale, although there are stratigraphic changes in the relative abundances of these two clay species through the section (Hayden Scott, 1992; Finkelstein, 1991). Secondly, scanning electron microscope (SEM) examinations of whole-rock specimens failed to reveal evidence of the large (>10 μm) euhedral chlorite grains that suggest in-situ formation. Clay mineralogists still disagree as to whether chlorite is an authigenic mineral at all; the chlorite appears to be detrital even in the drill cores from Gulf Coast wells (Weaver, 1989). Lastly, Rock-Eval data run on shale samples from all three sections show a T_{max} below the oil-generating window (Table 2). Translated into down-hole temperatures, this indicates a burial temperature of <75°C. Accordingly, based upon available evidence, we interpret the chlorite at the MV section as predominantly of detrital origin.

Kaolinite can also have a post-depositional origin; when pyrite and other sulfide minerals weather, they produce acidic solutions that can hydrolyze clay minerals into Al-rich varieties (Pollastro, 1985). Such processes are common in black shale facies. Again, we found little evidence for this process in the clay minerals of the samples examined. SEM examination of the mudstone did not reveal the distinctive books of kaolinite that characterize post-depositional growth of this mineral under acidic conditions. In addition, pyrite is only a minor constituent of the fine-grained sediments and black shales of the sort deposited in highly anoxic environments do not occur in the study interval. The MV section, as discussed below, has fewer indices of anoxia than either the LP or the RC section, so the weathering of sulfide minerals should not have a pervasive effect on the sedimentary mineralogy.

TABLE 2 - ROCK-EVAL DATA[2]

TOC and Rock-Eval Analyses from RC, MV, and LP								
Sample	**Tmax**	**Reliable?**	**S1**	**S2**	**S3**	**TOC**	**HI**	**OI**
BC-19	456	no	0.02	0.07	0.25	0.04	175	625
BC-16	439	no	0.05	0.06	0.57	0.06	100	950
BC-11	**433**	**yes**	0.07	0.8	0.88	0.44	182	200
BC-9	386	no	0.03	0.07	0.43	0.1	70	430
BC-5	417	no	0.07	0.25	0.73	0.14	179	521
BC-1	430	no	0.05	0.28	1.01	0.26	108	388
MV-29.8	**434**	**yes**	0.38	3.57	1.51	1.89	189	80
MV-25.2	439	no	0.01	0.04	0.41	0.08	50	513
MV-24.7	483	no	0.01	0.05	0.43	0.04	125	1075
MV-23.6	437	no	0.02	0.13	0.53	0.28	46	189
MV-22.7	503	no	0.02	0.12	0.36	0.21	57	171
MV-20.5	408	no	0.01	0.03	0.25	0.15	20	167
MV-18.5	501	no	0.01	0.07	0.26	0.26	27	100
LP-22	**430**	**yes**	0.02	2.47	0.76	1.11	223	68
LP-17	466	no	0.01	0.1	0.71	0.2	50	355
LP-14	**432**	**yes**	0.02	3.2	2.86	2.16	148	132
LP-12	**435**	**yes**	0.04	1.71	3.57	1.44	119	248
LP-9	436	no	0.01	0.42	1.75	0.75	56	233
LP-7	426	no	0.04	0.13	1.78	0.38	34	468

[2] Rock-Eval data from selected samples from the LP, MV, and RC sections. Tmax = temperature of maximum rate of evolution of S2 hydrocarbons (Tmax values are not necessarily reliable with S2<0.50); S1 = free, thermally extractable hydrocarbons; S2 = hydrocarbons from the cracking of kerogen and high molecular weight free hydrocarbons which did not vaporize in the S1 peak; S3 = organic carbon dioxide; TOC = total organic carbon (weight percent of carbon); HI = hydrogen index; and OI = oxygen index. Note that the samples with "reliable" values all yield similar Tmax values (other samples yield unreliable Tmax values due to low S2 values).

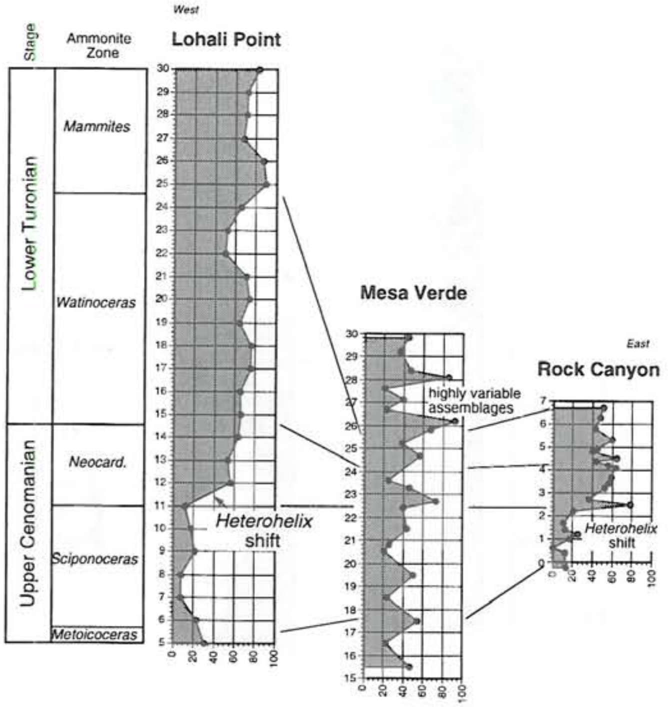

FIG. 9—Percent of the planktic foraminiferal genus *Heterohelix* at all three sites. Stratigraphic position of samples from the Mancos Shale at the LP and MV sections corresponds to meters above the Dakota Sandstone. At the RC section the sample position corresponds to meters above the base of the Bridge Creek Limestone Member of the Greenhorn Formation (lowest sample is from the uppermost Hartland Shale Member). Note the *Heterohelix* shift at the base of the *Neocardioceras* zone of sections at LP and RC and the highly variable populations at MV.

The foregoing considerations compel us to conclude that the clay minerals in the Cenomanian-Turonian interval of the three sections are largely detrital, and reflect conditions in the source areas and at the depositional sites in the seaway. Therefore, the stratigraphic changes in relative abundances record ancient fluctuations in the physical and chemical environment of the Western Interior Sea. The increasing abundance of illite-smectite beginning the base of the LP section is consistent with transgression of the sea during late Cenomanian time. Normally, illite-smectite consists of smaller particles and is deposited farther from shore than either kaolinite or discrete illite (Gibbs, 1977; Chamley, 1989). A decrease in kaolinite at LP is concomitant with transgression. A decrease in kaolinite also occurs at MV and RC, but in a more subdued fashion because those sections were located farther offshore ("A" in Figs. 8 and 10).

The "spikes" of kaolinite found in the *Sciponoceras/Neocardioceras* biozone transition, particularly in the sections at RC and LP, are a curious phenomenon ("B" and "C" in Figs. 8 and 10). One interpretation is that influx of kaolinite resulted from a short regressive event within this transgressive interval. We do not favor this interpretation because the spikes are more pronounced in the deepest, most distal environment. However, kaolinite production is enhanced during warm, wet climatic episodes and such conditions could also cause the kaolinite increase.

But why the greater effect at the most distal site? One possibility is that the sediment source was not from the Sevier orogenic belt west and northwest of the study sites. Instead, perhaps the kaolinite influx was derived from the south or southeast. The Sevier belt certainly supplied detritus, but the rapid rate of uplift and the short transport distances to the seaway may have delivered less weathered material to the depositional sites. In contrast, a warm, humid climate in less tectonically active southerly parts of the continent may have produced a greater abundance of kaolinite in the soils.

Increasing kaolinite abundance in the lower Turonian part of the LP section also corroborates this idea (Leckie et al., 1991). There, the supply of kaolinite increased despite a widening of the sea and deepening of the water column through the time of peak transgression in the early Turonian. The kaolinite presumably came from a deeply weathered source. Tethyan waters could have transported these clays into southern parts of the seaway. The section at MV shows some of these influences, but the very different clay assemblages suggest an alternate source area.

Comparisons Among the Sections

The LP and MV sections both show a dramatic decline in the proportion of planktics to total foraminifera in the uppermost Cenomanian *Neocardioceras* zone (Fig. 5). Such a trend may record a decline in planktic abundances, perhaps due to reduced salinities in the upper water column or to dissolution of the thinner-walled planktics. Conversely, the signal could reflect greater abundances of benthics due to improved conditions at the seafloor or an influx of food.

The last scenario is favored for the following reasons: (1) the clay mineralogy is very different in the sections at MV and LP, but if both sections had experienced the same low salinity event, then a similar spike in detrital clay minerals would be expected (Fig. 8); (2) the striking similarities in benthic assemblages at LP and MV (Fig. 7) suggest that both sections experi-

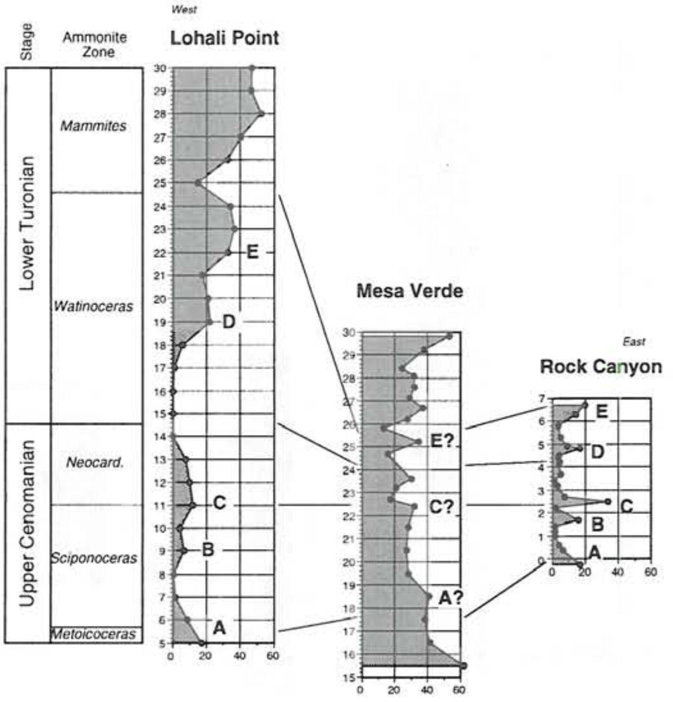

FIG. 10— Changes in percent kaolinite at the LP, MV, and RC sections. Letters A-E illustrate points of correlation between the LP and RC sections. Note the doublet kaolinite peaks in the *Sciponoceras-Neocardioceras* zone transition at LP and RC (peaks B and C). Refer to Fig. 9 caption for additional explanation.

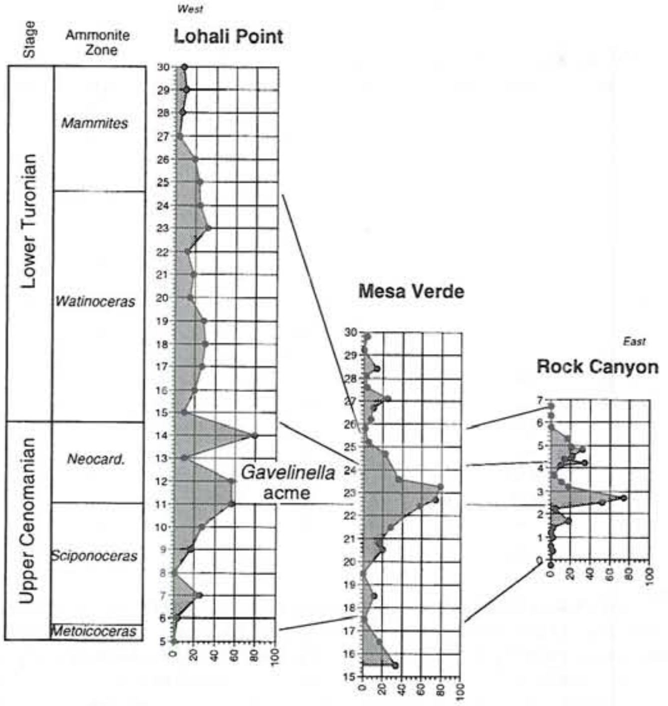

FIG. 11—Changes in percent *Gavelinella dakotensis* through the Cenomanian-Turonian boundary interval at the LP, MV, and RC sections. Note the acme of this taxon in the *Neocardioceras* zone of all three sites. Refer to Fig. 9 caption for additional explanation.

enced similar environmental conditions at the seafloor; (3) there is no evidence for differential dissolution of planktic tests; and (4) a relatively short-lived acme of *Gavelinella dakotensis*, a presumed epifaunal taxon, at all three localities (Fig. 11) may record an opportunistic response to a greater flux of organic matter to the seafloor from either terrestrial or marine sources (e.g., Gooday, 1993; Thomas and Gooday, 1996; West et al., this volume).

In further support of this hypothesis, Elder (1987, 1991) documented a dramatic increase in the detrital feeding gastropod *Drepanochilus* through the middle and upper parts of the *Neocardioceras* zone in sections from northeastern Arizona and south-central Utah. The highest organic carbon contents also occur in the *Neocardioceras* zone, which suggests that the availability of abundant organic matter may have stimulated growth of benthic foraminiferal populations.

At all three localities, the acme of *Gavelinella dakotensis* in the *Neocardioceras* zone is followed by a diachronous but abrupt change in structure of benthic foraminiferal populations (Figs. 7, 12). Assemblages characterized by high dominance of *Neobulimina albertensis*, an infaunal taxon, and very low diversity first appear in the lower part of the *Neocardioceras* zone at RC, in the middle part of that zone at MV, and at the top of the zone, which marks the Cenomanian-Turonian boundary, at LP. This appears to be a widespread event that cuts across paleodepths and facies. Elder (1990, 1991) also notes a diachronous change in molluscan faunas from the west-central part of the seaway toward the western margin during *Neocardioceras* time. Diverse to moderately diverse mixed infaunal-epifaunal biofacies of the *Sciponoceras* zone are replaced by depauperate, mixed infaunal-epifaunal, inoceramid-*Pycnodonte*, or other inoceramid-dominated biofacies. By early *Watinoceras* time (earliest Turonian), a more widespread and ubiquitous mix of inoceramid-*Entolium*, inoceramid-*Pseudoperna*, inoceramid-*Phelopteria*, or moderately diverse epifaunal biofacies lacking infaunal components characterized the southwestern and central parts of the seaway. The last stronghold for mixed infaunal and epifaunal molluscan communities through the Cenomanian-Turonian boundary interval was the clay-dominated western facies (Elder, 1990, 1991).

Why are the benthic foraminiferal assemblages from the LP section so similar to those at MV, while the clay mineralogy and stratigraphic trends in the planktic assemblages at LP are more similar to those at RC? If the water column were stratified by water masses from different sources, then this transect of sites across the southwestern side of the Greenhorn Sea records the character of those water masses where they met, interacted, and changed with rising sea level.

Upper Water Column

The *Heterohelix* shift coincides with a sharp depletion in $\delta^{18}O_{carb}$ recorded in the lower Bridge Creek Member of the Greenhorn Formation near Pueblo, Colorado (Pratt, 1985) (Fig. 2). Whole-rock oxygen isotope values drop from a high of about $-3.5^{o}/_{oo}$ in a limestone bed near the top of the *Sciponoceras* zone to about $-8.0\ ^{o}/_{oo}$ in a calcareous mudstone at the base of the *Neocardioceras* zone (Pratt, 1985; Pratt et al., 1993). Light values ($<-7\ ^{o}/_{oo}$) characterize all lithologies through the uppermost Cenomanian. The $\delta^{18}O_{carb}$ values actually fall to nearly $-9\ ^{o}/_{oo}$ in marly shales near the Cenomanian-Turonian boundary. Limestones through the basal Turonian show gradually heavier values (-7 to $-5\ ^{o}/_{oo}$), whereas the dark marlstones and marly shales remain markedly depleted but gradually become heavier (-8 to $-6.5\ ^{o}/_{oo}$) (Pratt et al., 1993).

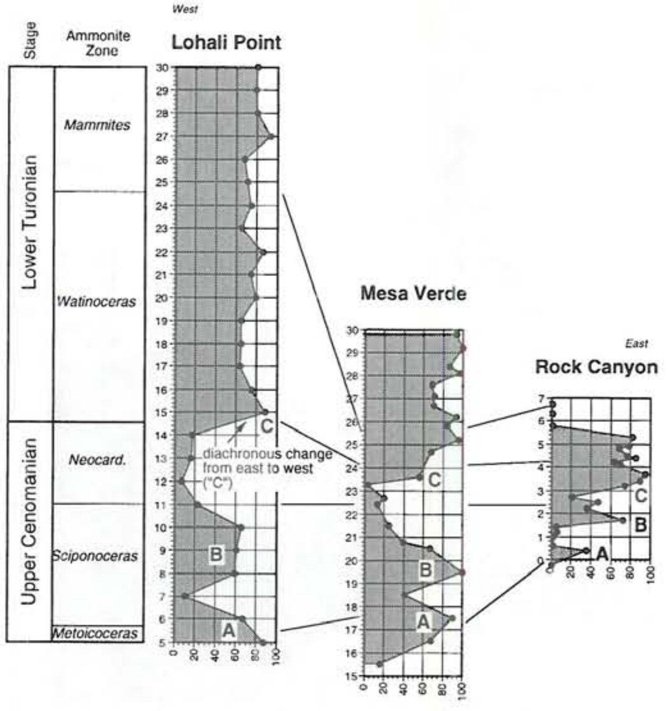

FIG. 12—Changes in percent *Neobulimina* (mostly *N. albertensis*) through the Cenomanian-Turonian boundary interval at the LP, MV, and RC sections. Note the abrupt but diachronous change to *Neobulimina* dominance through the uppermost Cenomanian *Neocardioceras* zone. Refer to Fig. 9 caption for additional explanation.

The abruptness and magnitude of the negative $\delta^{18}O_{carb}$ shift at the base of the *Neocardioceras* zone suggests that this signal records a major influx of fresh water to the seaway from riverine sources during latest Cenomanian time (Pratt, 1984, 1985; Barron et al., 1985; Pratt et al., 1993). This interpretation is supported by the coincident increase in discrete illite and kaolinite in a core from near Pueblo, Colorado (Pratt, 1984) and in our LP and RC sections (Fig. 8). In addition, an increase in apparent sedimentation rates in the *Neocardioceras* and lower *Watinoceras* zones in the central part of the seaway also supports Pratt's model, which points to increased runoff and fluvial input as the cause of the uppermost Cenomanian $\delta^{18}O$ depletion interval (Elder, 1985).

If we accept the hypothesis that LP and RC were influenced by the same southern surface water mass based on planktic foraminiferal assemblages and the same detrital source based on clay mineralogy, then the fresh water influx responsible for the negative excursion in $\delta^{18}O$ and detrital clay spike in the uppermost Cenomanian must have originated along the southeastern or southwestern part of the seaway and not in the Sevier orogenic belt of Utah-Nevada-Idaho-Wyoming. There is no doubt that the Sevier orogenic belt was an important primary source of riverine and clastic sediment input to the sea. But at times, particularly during transgressive pulses, clays across the southern part of the seaway may have come from additional sources. This interpretation is supported, in part, by the greater increase in illite + kaolinite in the basal *Neocardioceras* zone at RC as compared with the more proximal LP section (Fig. 8). Additional evidence comes from the distribution of macrofossils. For example, Elder (1985) suggested that the paucity of ammonites - including limestone - in the *Neocardioceras* zone at RC is perhaps related to the exclusion of ammonite larvae due to a stable subsaline surface layer on the sea. The few ammonites that occur at RC indicate warm southern affinities (Elder, 1985). However, ammonites are more common in this zone along the western margin (Elder, 1987, 1991; Kirkland, 1991). These observations raise further questions about the Sevier orogenic belt as the source of freshwater and detrital clays at LP.

Simulations of Cenomanian-Turonian climate were run to test if orbitally-induced insolation forcing, caused cyclic changes in precipitation intensity, as well as to test for the existence of high precipitation belts in or adjacent to the Western Interior Sea, which is required by the dilution model (e.g., Barron et al., 1985; Glancy et al., 1986, 1993). These models consistently indicate heavy winter precipitation along the northern continental margin of the Tethys Sea. This area includes the southern part of the Western Interior Sea and adjacent land masses. Glancy et al. (1986) suggest that the cyclonic rotation of winter storm tracks may have moved warm, moist air masses up the eastern slopes of the Sevier highlands, bringing greater precipitation and runoff under "maximum" forcing. Parrish (1993) also suggests that if the Sevier orogenic belt was high enough, it may have resulted in a low-pressure cell capable of drawing in moisture from the south, in spite of the possibility that the highlands may have also provided a rain shadow to zonal air flow.

The climate models do suggest that orbitally-driven changes in precipitation intensity may have influenced sedimentation patterns within the Western Interior Sea. In addition, the models also suggest that highlands of northern Mexico and/or southern Arizona, as well as the southern stable cratonic lowlands of the United States (e.g., Sohl et al., 1991) may have periodically experienced heavy winter rainfall because of its proximity to the northern margin of Tethys.

The preponderance of illite and the significant amount of chlorite at MV suggest that the clay mineral assemblages originated from a very different source area, one that was cooler and more temperate. We envision a circulation pattern in the Western Interior Sea that brought detritus from the Sevier orogenic belt and more northerly sources (Frontier Delta?) to sites in the vicinity of MV, while southerly sources supplied at least some of the muds deposited near LP and RC during the time of the Cenomanian-Turonian boundary. We propose that a broad offshore bathymetric high (forebulge?) separating the foredeep shelf from the deep central axis of the seaway bifurcated warm water currents (Fig. 13). Additional evidence of a bifurcating southern tongue comes from an independent analysis of the trace-element geochemistry of uppermost Cenomanian strata (Orth et al., 1993) (Fig. 14). Their elemental abundance data point toward a sediment source to the south and east.

Also noteworthy is an increase in kaolinite and illite in the upper part of the *Sciponoceras* zone at LP and RC that is associated with a brief but dramatic change in benthic foraminiferal population structure at RC (1.7 m) followed by brief recovery at the top of the zone (2.2 m). This pattern is a precursor to a second, larger detrital clay spike, *Heterohelix* shift, and negative oxygen isotope shift in the basal *Neocardioceras* zone (Figs. 2, 15). A similar pattern of biotic perturbation and subsequent recovery of molluscan communities occurred in the section at RC just prior to a faunal turnover and immigration event at the *Sciponoceras/Neocardioceras* zonal boundary (Elder, 1991) (Fig. 2). This upper *Sciponoceras* zone molluscan perturbation event is common at many localities (Elder, 1985). However, the clay

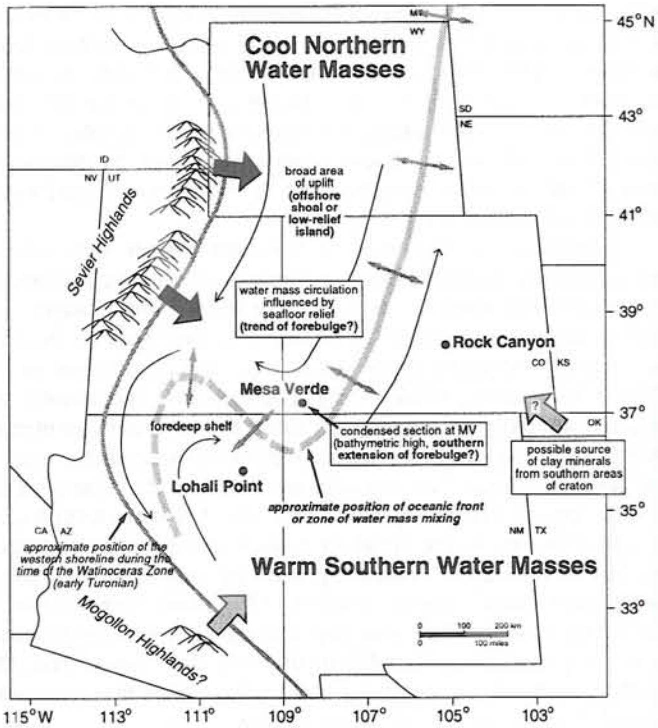

FIG. 13—Oceanographic interpretation of the Greenhorn Sea during latest Cenomanian-earliest Turonian time showing possible sources of clay minerals and a hypothetical circulation pattern that accounts for the temporal and spatial changes in clay mineralogy, and planktic and benthic foraminiferal paleobiogeography. Note the position of the Mesa Verde section along a presumed north-south trending bathymetric high, or forebulge. We suggest that this forebulge influenced the circulation of upper water masses on the western side of the seaway (thin arrows).

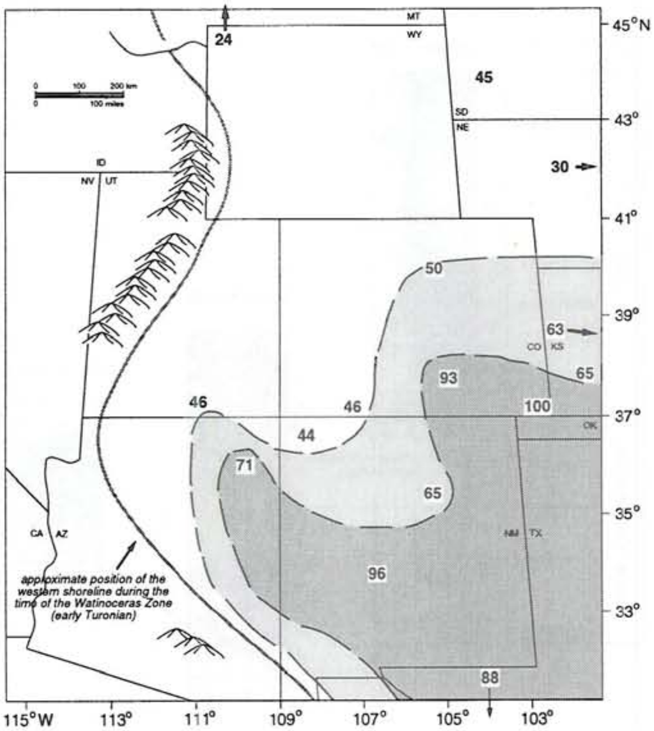

FIG. 14—Trace element concentration (Sc + Cr + Co + Ir) in upper Cenomanian rocks from across the Western Interior Sea (data from Orth et al., 1993). Values represent averages of the four elements through the *Sciponoceras* and lower *Neocardioceras* zones. These values appear to be bifurcated in the area around Mesa Verde.

spike and faunal events in the upper *Sciponoceras* zone are not associated with significant $\delta^{18}O$ depletion or with any change in planktic foraminiferal population structure at either LP or RC.

There appears to be a compelling, albeit complex, causal relationship between the *Heterohelix* shift, the $\delta^{18}O$ depletion event, and the clay-mineral changes. At the RC section in the center of the seaway, the *Heterohelix* shift coincides with a major influx of illite and kaolinite (2.5 m), but to the southwest, at LP, the change in planktic foraminiferal populations (12 m) occurs shortly after the small but similar clay spike (11 m) (compare Figs. 9, 10). Did *Heterohelix* respond to increased riverine input and development of a subsaline cap, as suggested by the $\delta^{18}O$ signal and the influx in detrital clays, or did it respond to other environmental variables? Leckie et al. (1991) suggested that the *Heterohelix* shift was related to the expansion or incursion of an oxygen minimum zone (OMZ) into the seaway rather than to slightly lower salinity conditions in the near-surface waters. Their rationale for preferring the OMZ hypothesis was based on the observation that *Heterohelix* continues to dominate the planktic foraminiferal assemblages at both LP and RC *above* the Cenomanian-Turonian boundary despite oxygen isotope data from the seaway that indicates a return to more normal marine conditions *at* the boundary (Pratt and Threlkeld, 1984; Pratt, 1985). In addition, following the detrital clay spike in the basal *Neocardioceras* zone, both kaolinite and illite drop off sharply and remain depressed across the Cenomanian-Turonian boundary. If *Heterohelix* responded mostly to reduced salinities, then its abundance, too, might be expected to drop off in concert with the detrital clays (Fig. 15).

The development of a widespread OMZ near the time of the Cenomanian-Turonian boundary was proposed to account for the unique biotic, sedimentologic, and geochemical signatures preserved in numerous localities around the world (e.g., Berger and von Rad, 1972; Frush and Eicher, 1975; Schlanger and Jenkyns, 1976; Hilbrecht and Hoefs, 1986; Arthur et al., 1987; Schlanger et al., 1987; Jarvis et al., 1988; Thurow et al., 1992; Arthur and Sageman, 1994; Kaiho and Hasegawa, 1994). In fact, the development or incursion of an OMZ into the southern Western Interior Sea may have been amplified by haline stratification caused by the development of a widespread, subsaline cap that originated from the southern part of the seaway during the latest Cenomanian. This scenario would have further decreased benthic ventilation, particularly in the deep central axis. The freshwater input at this time may be related to a fourth- or fifth-order sea level events as suggested by molluscan and dinoflagellate data (Elder, 1991; Li and Habib, 1996).

We propose that the influence and environmental impact of an OMZ expansion or incursion was diachronous from the central axis of the seaway toward the western margin during the latest Cenomanian. This hypothesis is supported by observed diachroneity in fossil assemblages and extinction events: (1) the *Heterohelix* shift relative to the clay influx noted above; (2) extinction of *Rotalipora* in the Western Interior Sea (Leckie, 1985); and (3) the observed diachroneity of biotic changes in both benthic foraminiferal and molluscan assemblages during *Neocardioceras* time (Elder, 1990, 1991).

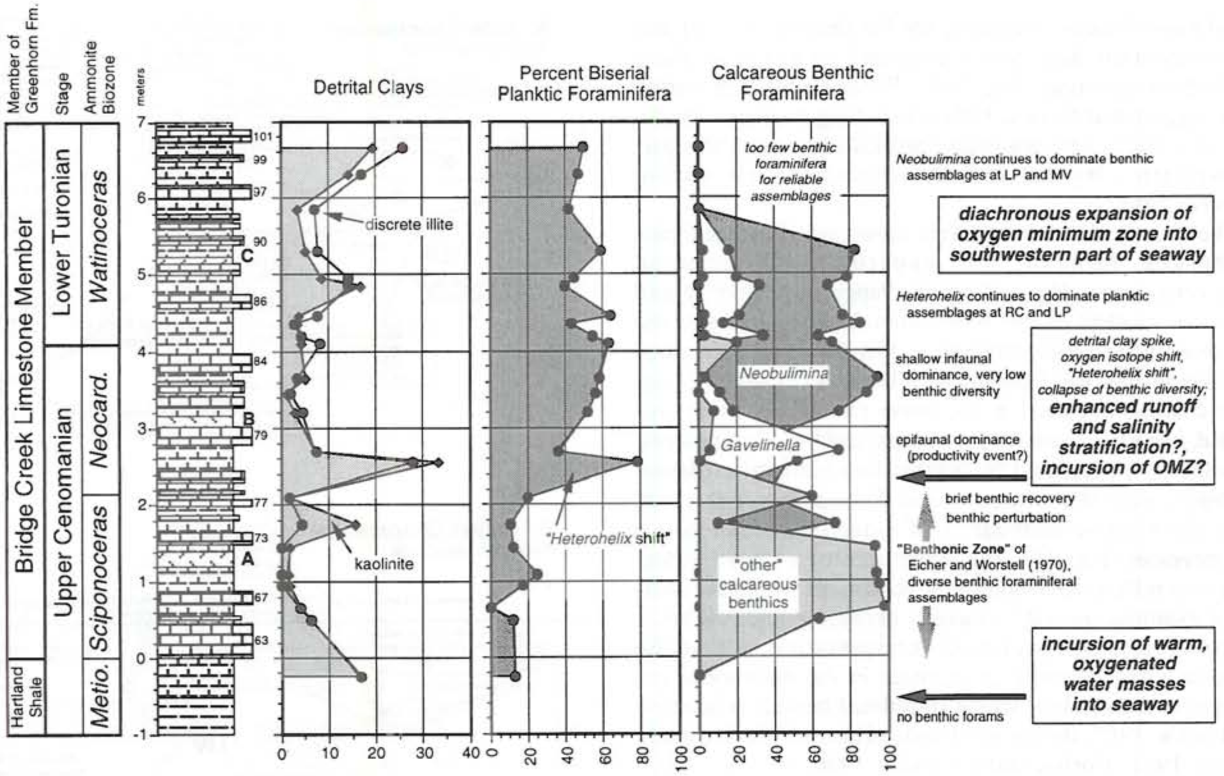

FIG. 15—Percentages of detrital clays, biserial planktic foraminifera, and calcareous benthic foraminifera through the Cenomanian-Turonian boundary interval at the Rock Canyon section, and interpretations of the changes observed in the clay mineralogy and in the foraminiferal assemblages.

Break-up of the subsaline cap, weakening of haline stratification, and a slight reduction in the intensity of the OMZ at the base of the *Watinoceras* zone may be due in part to a major transgressive pulse noted by Elder (1991). Such a hypothesis is supported by an abrupt increase in carbonate content at sections in the western clay-dominated facies and by a dramatic influx of *Mytiloides*, which is a turbidity-sensitive, epifaunal bivalve, at the base of the Turonian (Elder, 1985, 1987, 1991). However, we propose that the impact of an OMZ was still widely felt across southern parts of the seaway through the time of peak transgression in the late early Turonian (*Mammites* zone) based on: (1) the persistence of *Heterohelix* dominance through this interval (Leckie et al., 1991; West et al., this volume); (2) the low diversity assemblages of benthic foraminifera dominated by *Neobulimina* (Eicher and Worstell, 1970; Frush and Eicher, 1975; Eicher and Diner, 1985; Leckie, 1985; West et al., this volume); and (3) the widespread, depauperate communities dominated by epifaunal bivalves, which suggest low benthic oxygen conditions that limited bioturbation and benthic turbidity (Kauffman, 1977, 1984; Elder, 1985, 1987, 1990, 1991; Elder and Kirkland, 1985; Kirkland, 1991).

Lower Water Column

Are the similarities in upper Cenomanian benthic foraminiferal assemblages of LP and MV due to similar water depths, depositional facies, water masses bathing the seafloor in this part of the seaway, or a combination of these variables? And, why do all three sections show strikingly similar patterns of change in the uppermost Cenomanian and basal Turonian (Figs. 11, 12)? It is likely that the benthic biota records a complex history involving not only different water masses and mixing of water masses, including the possible incursion of an oxygen minimum zone as discussed above, but also changes in relative sea level, surface water productivity, and paleobathymetry. For example, transgressive pulses at the bases of the *Sciponoceras*, *Neocardioceras*, and *Watinoceras* zones changed substrate firmness and composition, benthic oxygen levels, and clastic sedimentation rates, all of which influenced the nature of the benthic molluscan communities (Elder, 1991).

During late Cenomanian time (*Sciponoceras* zone), the central axis of the seaway was dominated by diverse, calcareous benthic foraminiferal taxa with strong southern affinities (Eicher and Worstell, 1970; Frush and Eicher, 1975; Eicher and Diner, 1985; Leckie, 1985) (Fig. 7, Table 1). These assemblages included a mix of probable infaunal and epifaunal taxa. Although the sections at LP and MV lack this rich benthic microbiota, species richness is greatest in both sections in the *Sciponoceras* zone, and both include infaunal and epifaunal taxa. In addition, the LP and MV sections contain northern agglutinated taxa as well as sporadic abundances of miliolid benthic foraminifera (*Quinqueloculina moremani*), which may indicate the influence of warm, normal salinity waters. In the western, clay-dominated sections of northeastern Arizona and southern Utah, diverse benthic macrofossils occur in the *Sciponoceras* zone, including stenotopic bivalves, gastropods, and echinoderms, many of which are restricted to this zone (Elder, 1991; Kirkland, 1991). The composition of the molluscan and benthic foraminiferal assemblages indicate mixed biotas of southern and northern affinities during *Sciponoceras* time, but also normal salinity and well-oxygenated conditions at the seafloor. It seems possible that the lower part of the water column in this relatively shallow region of the Colorado Plateau, including the foredeep shelf (LP) and offshore bathymetric high (MV), consisted of mixed northern and southern water masses (Fig. 16A). Based on the rich marine biota, we suggest that there is little compelling evidence for the existence of a thick, well-developed brackish cap along this part of the western clay-dominated lithofacies belt during *Sciponoceras* time.

The benthos at all three localities shows significant and parallel changes through the uppermost Cenomanian *Neocardioceras* zone. The pattern of change from diverse and heterogeneous molluscan and foraminiferal communities of the *Sciponoceras* zone to high dominance, low diversity, ubiquitous communities across the Cenomanian-Turonian boundary suggests that changes occurred in the lower part of the water column related, in part, to benthic oxygenation and trophic resources. This hypothesis is supported by benthic foraminiferal evidence. For example, near the *Sciponoceras-Neocardioceras* zonal boundary the relative abundance of *Gavelinella dakotensis* abruptly increases (Fig. 11). The dominance of a single epifaunal taxon may reflect deterioration of conditions within the sediments -for example, anoxia? - thereby excluding infaunal taxa.

Alternatively, our preferred interpretation is that the "*Gavelinella* acme" records an increase in the delivery of organic matter to the seafloor which stimulated benthic productivity (e.g., Leckie, 1987; Berger and Diester-Haass, 1988; Herguera and Berger, 1991; Corliss and Emerson, 1990; Gooday, 1993; Jorissen et al., 1995; Thomas and Gooday, 1996; West et al., this volume) (Figs. 7, 11). The negative $\delta^{18}O_{carb}$ excursion and kaolinite + illite influx at the base of the *Neocardioceras* zone are attributed to a major influx of freshwater into the seaway. Such an event presumably would also be accompanied by an influx of dissolved nutrients and terrestrial organic matter. We documented a major increase in the proportion of benthic foraminifera to total foraminifera (% planktics) in the *Neocardioceras* zone in the sections at LP and MV (Fig. 5), which could be attributed to such a burst in benthic productivity rather than to increased dissolution of planktics. The increased supply of organic matter to the benthos could have been supplied by an influx of terrestrial material or as a rain of marine organic matter from increased surface water productivity (Fig. 16B). The dramatic increase in the abundances of the detritus-feeding gastropod *Drepanochilus* in the northeastern Arizona and southern Utah sections during *Neocardioceras* time was attributed to an influx of terrestrial organic matter (Elder, 1991).

Benthic productivity in the central axis of the seaway may have been suppressed by very low oxygen conditions near the seafloor and within the sediments, despite a possible concurrent influx of organic matter during the time of the *Neocardioceras* zone. This hypothesis is supported by foraminiferal and molluscan data at the RC section (Eicher and Diner, 1985; Elder, 1985; Elder and Kirkland, 1985; Leckie, 1985). Oxygen-poor intermediate waters from the south may have gradually invaded the western margin as sea level rose.

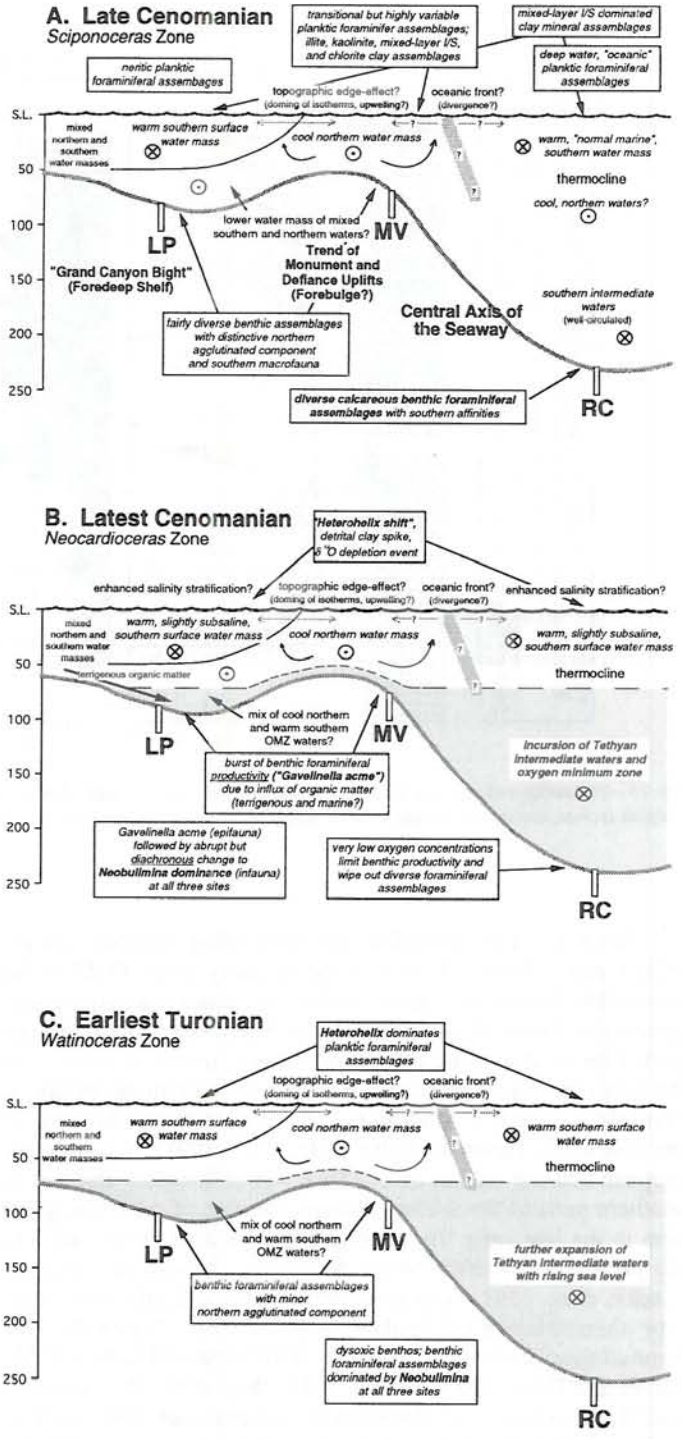

FIG. 16—Model of oceanographic conditions across the southwestern side of the Greenhorn Sea during the time of the Cenomanian-Turonian boundary; (A) late Cenomanian time (*Sciponoceras* zone), (B) latest Cenomanian time (*Neocardioceras* zone), and (C) earliest Turonian time (*Watinoceras* zone). Water depth is estimated in meters. Circles containing an "X" represent southern water masses flowing northward; circles containing a dot represent northern water masses flowing southward. Note rising sea level from panel A to C.

The incursion or expansion of an oxygen minimum zone may have been most intense during *Neocardioceras* time due to the development of a subsaline surface water mass as evidenced by the detrital clay spike. An influx of riverine waters from a southerly source caused enhanced stratification of the water column (Fig. 16B). The OMZ waters mixed with northern waters along the western margin and may have contributed to diachronous changes observed in the benthic assemblages of foraminifera and mollusks during the time of the *Neocardioceras* and *Watinoceras* zones (Fig. 16C). Weakening of a widespread OMZ during the early Turonian *Watinoceras* zone is suggested by the slight improvement in benthic conditions across a sea dominated by low diversity, depauperate communities (Eicher and Worstell, 1970; Elder, 1985, 1987, 1990, 1991; Elder and Kirkland, 1985; Eicher and Diner, 1985; Leckie, 1985; Leckie et al., 1991; Kirkland, 1991). A return to more normal marine surface waters is indicated by heavier whole-rock $\delta^{18}O_{carb}$ values during earliest Turonian time (Pratt, 1985; Pratt et al., 1993); one important consequence may have been a weakened salinity stratification of the water column.

Neobulimina, a presumed infaunal, calcareous benthic species, dominates the benthic assemblages (>60%) at all three sites in the *Watinoceras* zone (Figs. 7, 12). The diachronous change from *Gavelinella* dominance (epifaunal) to *Neobulimina* dominance across the Cenomanian-Turonian boundary may indicate subtle changes in the balance between organic matter flux to the seafloor, and dissolved oxygen at the sediment-water interface and in sediment pore waters. Some benthic foraminifera may have evolved symbiotic associations with chemoautotrophic bacteria (West, 1993), while others may be microaerophiles or facultative anaerobes (Bernhard, 1996), which potentially could confer an advantage for survival in dysoxic or anoxic environments. We suggest that *Neobulimina* was tolerant of the low oxygen conditions associated with the warm Tethyan water masses that filled the southern part of the seaway as sea level rose (West et al., this volume). The fact that macrofossil communities continue to be dominated by epifaunal bivalves to the near-exclusion of infaunal taxa in the *Watinoceras* zone suggests that the anoxic-dysoxic boundary in the sediments remained fairly close to the surface.

CONCLUSIONS

1. The late transgressive phase of the Cenomanian-Turonian Greenhorn Sea was characterized by the meeting and mixing of different water masses from northern and southern sources. Warm surface and intermediate waters moved rapidly northwards into the seaway during the late Cenomanian (*Sciponoceras* zone). These southern waters dominated the water column of the deep central axis of the seaway. Cooler, probably slightly less saline waters moved southwards along the western margin. An oceanographic front or mixing zone developed between the northward-flowing and southward-flowing water masses in the deeper, more stable central part of the seaway during the time of the Cenomanian-Turonian boundary. Benthic foraminiferal evidence suggests that warm southern waters overrode the cooler northern waters near Lohali Point in northeastern Arizona.

2. Bathymetric highs on the seafloor may have significantly influenced the circulation and mixing of water masses in the seaway. Seafloor relief on the present-day Colorado Plateau permitted the incursion of warm southern water masses along the foredeep shelf of northeastern Arizona and south-central Utah. Cooler northern waters dominated the broad, north-south bathymetric high (forebulge) to the east due to topographic mixing or surface water divergence and upwelling.

3. Clay mineralogy suggests detrital sources of mud in addition to the Sevier orogenic belt along the western side of the seaway. Significant riverine input from the southwest—possibly the Mogollon highlands or northern Mexico—or from the southeast—southern U.S. craton—may have occurred during periods of dark marlstone or calcareous shale deposition across the southern Greenhorn Sea (Fig. 13).

4. Changes in planktic and benthic foraminifera communities from the southwestern side of the Greenhorn Sea through the Cenomanian-Turonian boundary interval, particularly through the uppermost Cenomanian *Neocardioceras* zone, reflect a combination of oceanographic variables and continued rising sea level (Fig. 15). Influx of freshwater to the seaway during the latest Cenomanian may have originated from a southern or southeastern source, as indicated by clay mineral data, but was also accompanied by the incursion of oxygen-poor intermediate waters from Tethys.

This latter interpretation is supported by foraminiferal and molluscan data. Influx of terrestrial organic matter and enhanced primary productivity associated with water-mass mixing and the influx of detrital sediments and nutrients at this time resulted in a burst of benthic biomass on the shallower southwestern side of the seaway. A widespread but short-lived abundance maximum of epifaunal *Gavelinella dakotensis* is interpreted as a response to increased organic matter flux to the seafloor. Enhanced salinity stratification in the deeper central part of the seaway may have amplified the affects of an oxygen minimum zone thereby severely limiting benthic productivity there. The influence of the OMZ gradually infiltrated the lower part of the water column on the western side of the seaway based on the diachronous changes in benthic foraminiferal and molluscan assemblages through the *Neocardioceras* zone (Fig. 16). *Neobulimina albertensis*, an infaunal taxon that was probably tolerant of low-oxygen conditions, dominates the benthic foraminiferal assemblages at all three localities in the basal Turonian *Watinoceras* zone when warm oxygen-poor intermediate waters invaded the southern seaway with rising sea level.

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APPENDIX A. Raw data from Lohali Point (LP). Foraminiferal data represent numbers of specimens. Clay mineral data are expressed as percentage of total clays.

Sample	Meters Above Dakota Sandst.	Total Planktics	Total Benthics	Total Foraminifers	Biserial *Heterohelix*	Triserial *Guembelitria*	Trochospiral planktics	Planispiral planktics	Keeled planktics	*Neobulimina*	*Gavelinella*	Other Calcareous	Agglutinated	% Mixed-Layer Illite-Smectite	% Discrete Illite	% Kaolinite	% Chlorite
LP30	30 +/-	233	49	282	191	6	36	0	0	40	4	2	3	43.9	9.4	46.7	0.0
LP29	29 +/-	221	72	293	159	1	59	2	0	55	7	1	9	44.9	8.8	46.3	0.0
LP28	28 +/-	201	92	293	142	7	42	7	0	78	6	3	5	36.6	11.2	52.2	0.0
LP27	27 +/-	183	97	280	124	3	56	0	0	93	4	0	0	54.1	5.6	40.3	0.0
LP26	26 +/-	177	111	288	153	0	24	0	0	87	24	0	0	57.7	10.4	31.9	0.0
LP25	25 +/-	184	103	287	165	5	14	0	0	73	24	1	5	77.5	7.9	14.6	0.0
LP24	24 +/-	174	113	287	114	16	44	0	0	84	28	0	1	52.7	13.2	34.1	0.0
LP23	23 +/-	155	150	305	81	44	30	0	0	96	47	2	5	57.5	5.9	36.5	0.0
LP22	22 +/-	159	127	286	80	35	44	0	0	109	15	0	3	60.5	6.8	32.7	0.0
LP21	21 +/-	179	123	302	127	37	15	0	0	95	23	3	2	78.5	3.9	17.6	0.0
LP20	20 +/-	191	93	284	141	30	20	0	0	74	14	3	2	69.9	9.4	20.7	0.0
LP19	19 +/-	220	99	319	146	41	33	0	0	64	28	1	6	72.0	6.1	21.9	0.0
LP18	18 +/-	214	101	315	163	31	20	0	0	65	30	3	3	92.2	1.9	5.9	0.0
LP17	17 +/-	173	136	309	127	22	24	0	0	87	37	11	1	97.0	2.0	1.0	0.0
LP16	16 +/-	122	203	325	79	27	16	0	0	145	39	3	16	98.3	1.7	0.0	0.0
LP15	15 +/-	101	200	301	67	17	17	0	0	172	17	4	7	97.0	3.0	0.0	0.0
LP14	14 +/-	107	197	304	68	29	10	0	0	34	155	2	6	97.8	2.2	0.0	0.0
LP13	13 +/-	67	114	181	36	25	6	0	0	35	22	1	56	89.4	2.9	7.7	0.0
LP12	12 +/-	215	120	335	120	82	13	0	0	10	82	9	19	84.2	5.8	10.0	0.0
LP11	11 +/-	292	43	335	34	257	1	0	0	11	24	3	5	80.0	8.0	12.0	0.0
LP10	10 +/-	267	58	325	46	208	13	0	0	38	16	1	3	93.5	2.2	4.3	0.0
LP9	9 +/-	116	190	306	25	86	5	0	0	114	31	3	42	90.3	2.9	6.8	0.0
LP8	8 +/-	310	144	454	33	271	6	0	0	95	2	7	40	98.4	1.1	0.5	0.0
LP7	7 +/-	285	18	303	17	230	38	0	0	2	5	9	2	96.3	2.4	1.3	0.0
LP6	6 +/-	213	74	287	50	136	27	0	0	47	2	14	11	87.8	3.5	8.7	0.0
LP5	5 +/-	165	117	282	55	0	98	12	0	97	1	0	19	77.9	5.2	16.9	0.0
LP4	4 +/-	90	152	242	41	32	0	17	0	86	5	20	41	65.8	7.9	26.3	0.0
LP3	3 +/-	296	43	339	142	67	84	3	0	6	8	13	16	50.0	6.4	43.6	0.0
LP2	2 +/-	6	131	137	3	3	0	0	0	21	4	10	96	58.0	10.5	31.5	0.0
LP1	1 +/-	55	181	236	7	17	31	0	0	15	23	0	143	24.8	16.8	58.4	0.0

APPENDIX B. Raw data from Mesa Verde (MV). Foraminiferal data represent numbers of specimens. Clay mineral data are expressed as percentage of total clays.

Sample	Meters Above Dakota Sandst.	Total Planktics	Total Benthics	Total Foraminifers	Biserial *Heterohelix*	Triserial *Guembelitria*	Trochospiral planktics	Planispiral planktics	Keeled planktics	*Neobulimina*	*Gavelinella*	Other Calcareous	Agglutinated	% Mixed-Layer Illite-Smectite	% Discrete Illite	% Kaolinite	% Chlorite
MV29.8	29.8	227	56	285	90	0	137	0	0	55	3	0	0	9.1	29.1	53.0	8.8
MV29.2	29.2	201	15	216	133	0	68	0	0	15	0	0	0	12.5	40.6	38.0	8.9
MV28.4	28.4	259	44	303	119	1	139	0	0	36	8	0	0	35.6	33.9	24.1	6.4
MV28.1	28.1	313	46	359	263	0	50	0	0	45	1	0	0	28.6	34.3	31.6	5.5
MV27.6	27.6	274	17	291	53	0	221	0	0	16	1	0	0	30.0	33.3	32.0	4.7
MV27.1	27.1	197	48	245	90	3	104	0	0	35	12	0	1	21.4	42.9	28.9	6.8
MV26.7	26.7	242	12	254	48	3	191	0	0	11	1	0	0	35.2	24.1	36.8	3.9
MV26.2	26.2	130	223	353	123	0	7	0	0	213	9	1	0	30.4	37.0	27.5	5.1
MV25.8	25.8	208	153	361	142	2	64	0	0	141	3	7	2	40.0	40.0	12.8	7.2
MV25.2	25.2	187	157	344	78	7	101	1	0	142	15	0	0	25.0	33.3	34.4	7.3
MV24.7	24.7	157	115	272	84	30	43	0	0	95	20	0	0	60.0	20.0	15.6	4.4
MV23.6	23.6	88	206	294	10	32	44	2	0	141	50	12	3	15.8	50.0	29.8	4.4
MV23.2	23.2	13	276	289	5	5	3	0	0	23	218	16	19	38.1	38.1	20.4	3.4
MV22.7	22.7	239	111	350	102	115	22	0	0	32	75	1	3	32.5	42.5	17.4	7.6
MV22.4	22.4	88	244	332	25	17	43	3	0	106	104	18	16	15.6	40.6	32.0	11.8
MV21.5	21.5	106	237	343	24	58	21	3	0	148	37	1	51	17.5	42.5	27.9	12.1
MV20.8	20.8	206	89	295	25	156	19	6	0	78	1	0	10				
MV20.5	20.5	138	49	187	19	66	52	1	0	35	6	2	6	11.5	45.3	27.2	16.2
MV19.5	19.5	38	47	85	16	4	18	0	0	14	3	1	29	15.4	46.2	28.0	10.5
MV18.5	18.5	167	34	201	30	45	91	1	0	21	4	8	1	4.4	45.1	40.6	9.9
MV17.5	17.5	194	105	299	67	45	80	2	0	102	2	0	1	8.3	36.7	38.3	16.7
MV16.5	16.5	260	82	342	42	4	203	9	2	48	8	26	0	16.7	26.7	41.0	15.7
MV15.5	15.5	239	67	306	85	17	132	5	0	26	5	18	18	8.3	30.6	61.6	0.0
MV14.5	14.5	67	131	198	13	14	40	0	0	55	19	16	41	9.7	32.3	58.0	0.0
MV13.5	13.5	204	69	273	53	93	55	3	0	44	19	0	6	6.6	30.3	63.1	0.0
MV12.5	12.5	9	46	55	0	4	5	0	0	31	7	1	7	7.7	38.3	44.9	8.9

APPENDIX C. Raw data from Rock Canyon (RC). Foraminiferal data represent numbers of specimens. Clay mineral data are expressed as percentage of total clays. Additional benthic specimens were picked for samples BC1-BC16 (Leckie, 1985).

Sample	Meters Above base Brdg Crk	Total Planktics	Total Benthics	Total Foraminifers	Biserial *Heterohelix*	Triserial *Guembelitria*	Trochospiral planktics	Planispiral planktics	Keeled planktics	*Neobulimina*	*Gavelinella*	Other Calcareous	Agglutinated	Mixed-Layer Illite-Smectite	Discrete Illite	Kaolinite	Chlorite
BC22	6.65	333	1	334	167	1	165	0	0	1	0	0	0	54.8	25.8	19.4	0.0
BC21	6.30	207	0	207	98	0	109	0	0	0	0	0	0	68.6	17.1	14.3	0.0
BC20	5.85	306	0	306	128	0	178	0	0	0	0	0	0	89.3	7.1	3.6	0.0
BC19	5.30	288	45	333	170	10	108	0	0	37	7	1	0	87.2	7.7	5.1	0.0
BC18	4.95	284	50	334	128	0	153	3	0	39	10	1	0	77.1	14.3	8.6	0.0
BC17	4.85	248	53	301	100	0	148	0	0	36	17	0	0	69.0	14.3	16.7	0.0
BC16	4.45	449	172	621	288	2	159	0	0	261	71	12	0	88.5	7.7	3.8	0.0
BC15	4.35	381	68	449	164	2	215	0	0	279	42	6	0	93.8	2.8	3.4	0.0
BC14	4.20	411	94	505	225	4	182	0	0	191	102	11	0	91.5	4.2	4.3	0.0
BC13	4.10	345	74	419	217	7	120	1	0	239	34	68	0	87.5	8.3	4.2	0.0
BC12	3.65	502	13	515	291	0	210	1	0	363	12	8	0	91.5	3.4	5.1	0.0
BC11	3.45	512	48	560	284	5	222	1	0	275	33	2	0	96.9	2.1	1.0	0.0
BC10	3.20	485	57	542	249	6	230	0	0	257	63	28	0	92.5	4.5	3.0	0.0
BC9	2.70	335	22	357	122	0	211	2	0	52	195	17	0	85.4	7.3	7.3	0.0
BC8	2.55	522	53	575	404	1	106	11	0	140	154	4	0	38.9	27.8	33.3	0.0
BC7	2.10	299	27	326	61	3	226	9	0	114	12	186	0	96.5	1.7	1.8	0.0
BC6	1.75	429	8	437	45	1	374	9	0	156	38	22	0	80.3	4.2	15.5	0.0
BC5	1.45	593	17	610	73	13	456	51	0	15	7	279	0	98.3	0.0	1.7	0.0
BC4	1.10	323	95	418	81	28	180	33	1	20	0	297	0	98.7	0.0	1.3	0.0
BC3	0.95	421	166	587	70	31	262	58	0	7	5	296	2	98.3	0.0	1.7	0.0
BC2	0.65	477	22	499	1	2	428	46	0	6	1	335	0	92.1	3.9	4.0	0.0
BC1	0.50	663	72	735	80	24	519	36	4	118	5	211	0	87.5	6.2	6.3	0.0
HL1	-0.25	352	1	353	44	0	282	11	15	0	0	1	0	66.7	16.6	16.7	0.0

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