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**Dinoflagellate response to sea level change in the Cretaceous
Western Interior**

Li, Huan, Ph.D.

City University of New York, 1994

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A

**DINOFLAGELLATE RESPONSE TO SEA LEVEL CHANGE
IN THE CRETACEOUS WESTERN INTERIOR**

by

HUAN LI

A dissertation submitted to the Graduate
Faculty in Earth and Environmental Sciences
in partial fulfillment of the requirements
for the degree of Doctor of Philosophy, The
City University of New York

1994

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This manuscript has been read and accepted for the Graduate Faculty in Earth and Environmental Sciences in satisfaction of the dissertation requirement for the degree of Doctor of Philosophy.

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ABSTRACT

DINOFLAGELLATE RESPONSE TO SEA LEVEL CHANGE
IN THE CRETACEOUS WESTERN INTERIOR

by

Huan Li

Adviser: Professor Daniel Habib

Detailed palynological analysis of depositional environments in carbonaceous shales and marlstones in the Cenomanian-Turonian(C-T) interval of the Western Interior Cretaceous Seaway indicates that (1) dinoflagellate species diversity, (2) ratio of *Spiniferites* to *Cyclonephelium* (S/C ratio) and (3) depositional organic facies are collectively useful for distinguishing cycles of marine transgression and regression. The increase in number of dinoflagellate species and of the S/C ratio within a matrix of abundant amorphous debris is interpreted to have resulted from the expansion of an epeiric seaway. The marked drop in species diversity in the interval of mixed amorphous debris and detrital inertinite

adjacent the *Sciponoceras-Neocardioceras* zonal boundary and prior to or at the C-T boundary indicates two short periods of shoreline progradation. The S/C ratio increases in the offshore direction as well as in the transgressive intervals. The largest number of species occurs in the lower part of the transgressive intervals, and not in peak transgression. The successive decrease in species diversity from the early stage to late stage within the transgressive interval and into the next highstand interval is the result of shoreline progradation. The downlap surface (maximum starvation surface) which delimits the boundary between the transgressive and highstand intervals may be indicated by an inflection in the trend of decreasing species diversity. The position of the inflection point within the C-T transgressive sequence shows the same trend as the position of the condensed interval that shifts higher in the offshore direction.

The preservation of dinocysts in sediments is highly variable and closely corresponds to lithology, with best in calcareous shales and worst in limestones and marlstones. The concentration of dinocysts in the C-T sediments decreases semi-logarithmically with increasing percentages of carbonate fraction. The organic matter deposited in the C-T sediments is mostly marine amorphous fecal debris and there is little landplant detritus even within the interval of progradation. The amount of terrigenous inputs from the western highland may

not be as large as other studies showed. The bottom condition in the coastal and nearshore areas may have been oxygenated or sub-oxygenated during transportation and deposition, providing the potential for oxidation and degradation of landplant material.

ACKNOWLEDGEMENTS

The author would like to express his deep gratitude and appreciation to his adviser Dr. Dan Habib for his help, guidance and encouragement throughout this study. Many thanks are extended to Dr. William Elder (U.S. Geological Survey, Menlo Park, California) for his support in providing most samples studied and many invaluable suggestions. The author is grateful to the faculty and staff of the Geology Department at Queens College for their support and guidance. Thanks are also extended to Dr. David Thurber and Dr. John Chamberlain for critically reviewing the manuscript. The author would like to thank Dr. Erle Kauffman for spending time with the author during the field work. Thanks are also due to his fellow students for companionship through many years. The author especially indebted to his family and friends for their understanding, patience and encouragement. Without them, none of this would have been possible. This study was supported in part by a grant from the National Science Foundation at Queens College, ERA-8903522.

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INTRODUCTION

Marine sediments deposited in deep oceans, continental shelves and epicontinental seas during late Cenomanian and early Turonian time (Fig. 1) display anomalous lithological, geochemical and faunal characteristics which reflect global paleoceanographic and climatic changes. Compared with most of the Phanerozoic record, uncommonly high organic carbon content and a positive shift in the $\delta^{13}\text{C}$ in C-T marine sediments indicate that a global burial of organic carbon occurred during marine transgression in late Cenomanian-early Turonian time (e.g., Stein 1986; Herbin et al., 1987; Arthur et al., 1988; Farrimond et al., 1990; Gale et al., 1993). These, and other features, e.g. lack of benthonic faunas, have been interpreted as indicating that significant parts of the world ocean were intermittently oxygen deficient across the C-T boundary (Fig. 2). This episode was termed C-T Oceanic Anoxic Event (OAE) by Schlanger and Jenkyns (1976), Arthur and Schlanger (1979), and Jenkyns (1980).

The C-T boundary is marked worldwide by an obvious faunal extinction in ammonites, inoceramid bivalves, and planktonic and benthonic foraminifera. The dramatic change in marine biota is probably the result of sea level rise, rapidly

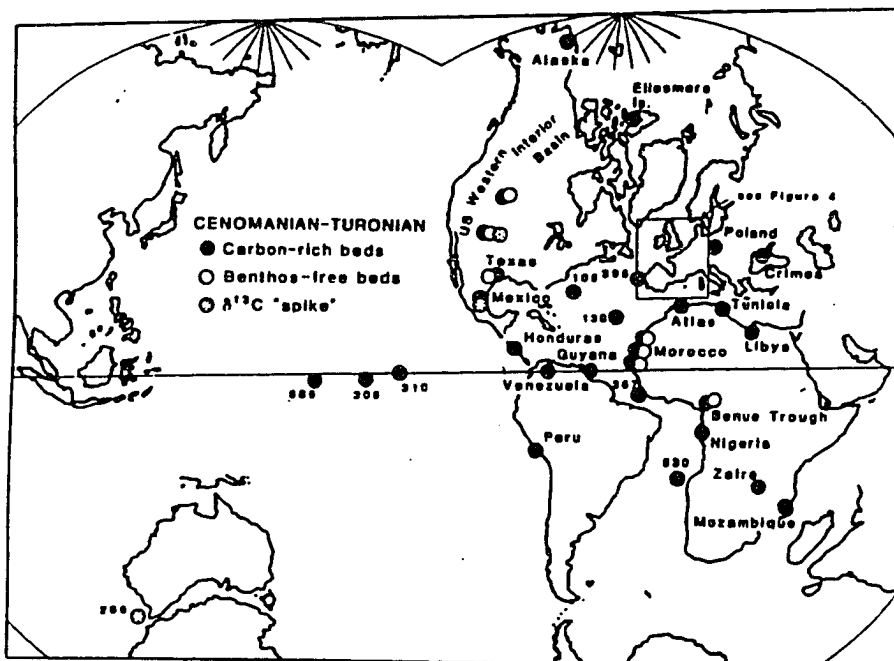


Fig. 1 Map of the world at 90 mybp showing locations of Cenomanian-Turonian sections (other than those in Europe). Deep Sea Drilling Project Sites are designated by number. After Schlanger et al. 1987.

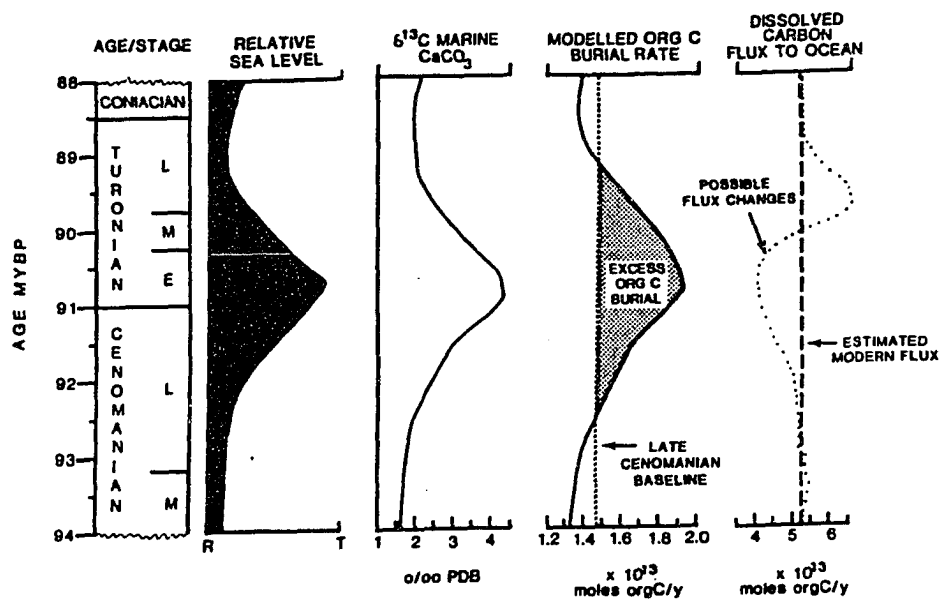


Fig. 2 Generalized diagram of Cenomanian-Turonian boundary eustatic and geochemical events. After Arthur et al. 1987.

changing oceanic circulation patterns, and local or global anoxic events which encroached in epicontinental seas (Kauffman 1977, 1984; Jarvis et al., 1988a; Elder 1987a, 1987b, 1989).

Since the level of oxicity/anoxicity within sediments and within the water column affects the amount and nature of the organic matter preserved in the rocks, the depositional environment can be characterized by determining the kind and amount of organic matter that accumulated and was preserved (Didyk et al., 1978). Modern black muds form in calm oxygen-depleted water, such as stagnant basins (fjords, Black Sea) (Roman 1974; Shimkus and Trimonis 1974; Traverse 1974), or below zones of high primary productivity in oxygenated waters (e.g. upwelling zones off Peru) (Suess et al., 1987). Because of different capacities to induce anoxia, the ratio between terrigenous and pelagic marine organic matter is important for interpreting the depositional environment (Tissot et al., 1979). Habib (1979, 1982, 1989) introduced a series of palynological facies, based on the abundance and character of palynomorphs and the nature of other carbonaceous materials in the organic matrix. These organic facies were used to interpret the extent of marine transgression, and the proximity to fluvial source(s). The palynofacies concept, along with sedimentological and geochemical evidence, has been used to interpret changes of sea level, oceanography and climate. The detailed study of dinoflagellates plays a very

important role in Habib's organic facies analysis and interpretation of depositional environments.

Among all known marine Cenomanian-Turonian boundary sections, those in the Cretaceous Western Interior seaway show the most completely exposed rock sequences (bioturbated carbonates to laminated, organic-rich shales) and the most complete fossil record (ammonites, bivalves, foraminifera, nannofossils). Detailed studies of high-resolution stratigraphy, globally and regionally precise correlation, and stable isotope geochemistry make it possible to interpret the eustatic, tectonic, sedimentologic, and ecologic-paleobiologic evolution of this large interior seaway (Kauffman 1977, 1984; Elder 1985, 1987b, 1989; Pratt 1984; Wright 1987). There are many studies demonstrating the usefulness of groups such as foraminifera, calcareous nannofossils and invertebrate macrofossils in the environmental analysis of the Western Interior basin, but relatively little attention has been paid to dinoflagellate cysts. It has been shown in other areas that the study of dinoflagellates can be useful for precise stratigraphic dating and correlation, and for interpretation of depositional environments, especially in regions depositing organic-rich sediments (e.g. Partridge 1976, Habib and Miller 1989, van Pelt 1990). The controversy over the origin of Cretaceous organic-rich sediments has brought about a general appreciation among the geological community that both geochemical and palynological studies of organic matter are an

essential prerequisite to any meaningful genetic interpretation of organic-rich sediments (e.g. Arthur 1979, Weissert 1981).

The purpose of this study is to examine differences in species diversity and specimen abundance in dinoflagellate assemblages and in organic facies within specific lithofacies regions to determine the effects that cyclic environmental changes, especially marine transgression-regression cycles, had on the distribution of dinoflagellate and other organic particles. Furthermore, these results will be used to reconcile currently-held ideas on the nature of the paleoceanography of the seaway during the Cenomanian-Turonian.

BACKGROUND STUDIES

Dinoflagellates and Depositional Organic Facies

1. Dinoflagellate Paleoecology

Dinoflagellates are opportunistic aquatic algae which occupy a wide variety of habitats ranging from the open ocean, to the continental margin, and to freshwater lakes (Habib and Miller 1989). A relatively small proportion of dinoflagellates can produce fossilizable cysts which are considered to be an adaptation to the unstable hydrographic regimes of shelf seas (Wall et al., 1977; Margalef et al., 1979). These cyst-producing (meroplanktonic) dinoflagellates are predominantly neritic (continental margin) so that there is a reasonable probability of the motile cell reaching the euphotic zone after excystment from a benthic cyst. Most oceanic dinoflagellates do not produce fossilizable cysts. The presence of dinocysts in oceanic sediments presumably is the result of transport either by surface currents or bottom currents which have traversed shelf areas, or by turbidity flows (Wall 1971; Wall et al., 1977; Dale 1976, 1983; Tyson 1987). The cyst assemblage does not reflect accurately either the qualitative or quantitative aspects of the motile community. This is even true in areas of relatively simple

hydrographic conditions for which plankton records have been maintained for several years (Wall et al., 1977; Reid 1978). Any ecological interpretation of cyst assemblages can be made if the following factors are taken into consideration: (1) taxonomically restricted cyst production(few species) (Dale 1976); (2) variable and largely undocumented cyst/theca abundance ratios (Reid and Harland 1977); (3) differential preservation (Dale 1976; Reid 1974; Harland 1977); (4) dispersion by currents (Wall et al., 1977); (5) sparse data on encystment triggering mechanisms (e.g., light, temperature); (6) other factors including processing techniques (Reid 1974; Dale 1976).

Despite the fact that the ecological significance of the dinoflagellate fossil record is limited in several respects, a relatively recent development in dinoflagellate micropaleontology has been the use of cysts as paleoenvironmental indicators. Dinoflagellate cysts can be used to indicate a general paleoenvironmental trend within a depositional basin by correlating changes in assemblage composition, species diversity and relative abundance to other independent criteria (e.g. other fossil groups, facies, geochemical, seismic, depositional sequences, etc.) because the cyst assemblages are ultimately related to environmental (hydrodynamic and climatic) factors and thus have paleoecological meaning in themselves (Wall et al., 1977). It is particularly significant to use distinctive associations or

particular taxa of dinocysts to infer paleoenvironments in those regions where there are no other available data.

Davies et al. (1982) reviewed the status of dinoflagellate paleoecology and classified the various techniques that have been used into four categories. They are: (1) relative abundance of dinoflagellates to other palynomorph types or plankton groups; (2) dinoflagellate species diversity and dominance; (3) relative abundance of species, groups of related species, or morphotypes within particular assemblages; (4) absolute abundance of dinoflagellate cysts per unit volume of sediment or water. The paleoecological application of all four parameters will be briefly discussed below.

Relative Abundance of Dinocysts

With respect to other microfossil groups, the relative abundance of dinocysts is a technique that has been applied to Quaternary (Reid 1975) as well as to older assemblages. The most commonly used ratio is that between dinoflagellate cysts and sporomorphs. The relative frequencies of the above fossil groups reflect changes in environmental conditions probably controlled by shoreline displacement (Manum 1976). Partridge (1976) correlated the occurrence of dinoflagellate cysts with shallow marine incursions within an overall non-marine depositional sequence in the lower Tertiary of Australian Gippsland Basin. Non-marine deposits in this area are characterized by distinctive sporomorph assemblages which lack

dinoflagellate fossils. Mebradu (1978) calculated numerous relative abundance ratios among four palynomorph categories (microplankton, microforaminiferal linings, sporomorphs, tasmanitids) for the Upper Jurassic of Dorset, England, and suggested that changes in the ratios were related to environmental fluctuation. He correlated the high relative abundance of microplankton (predominantly dinoflagellate cysts) to transgressive, or deep water, depositional phases in the basin. Changes in the relative abundance of dinoflagellate cysts have also been linked to fluctuations in organic carbon and phosphate content in sediments from the mid-Cretaceous of the Paris Basin (Fauconnier and Slansky 1978), and in the Paleocene and Eocene of the Gafsa Basin, Tunisia (Fauconnier and Slansky 1978, 1980). McMinn's study (1991) of estuaries on the central coast of New South Wales, Australia shows that the relative abundance of dinocysts is consistent with change in specimen abundance, with lower values closer to the shores and higher values in the deeper area.

Species Diversity and Dominance

Species diversity (species richness and species dominance) is a variable with potential utility in dinoflagellate paleoecology. Species diversity is measured by the number of species in a sample. In general, dinoflagellate species diversity changes in an onshore to offshore direction. Offshore shallow shelf assemblages are richer on the average

than nearer-shore assemblages (Wall et al., 1977; Dale 1983). Like all other organisms, the number of dinoflagellate species present is proportional to the size of area. Thus, an expanded continental shelf formed as the result of marine transgression provides more habitats and leads to increased species abundance (MacArthur and Wilson 1967). The generally good agreement between dinoflagellate species diversity and the cycles of relative sea-level change has been proposed by Habib and Miller (1989). The other factors probably affecting dinoflagellate species diversity include nutrient supply (related mostly to upwelling), oceanic circulation, seasonal fluctuation and salinity change. Studies for recent phytoplankton and ancient fossils indicate that coastal and estuarine assemblages tend to consist of fewer species, due mainly to reduced salinity or unstable temperature fluctuation (Hurlbert 1963; Williams and Sarjeant 1967).

The significant increase in dinoflagellate species diversity associated with the amorphous debris facies is considered to be palynological evidence of marine transgression (Habib and Miller 1989; van Pelt 1990). Partridge (1976) suggested that eustatic sea level rise caused the rapid increase in the number of dinoflagellate species in sediments, and that the occurrence of these dinoflagellate incursions can be used to interpret eustatic sea level changes in other areas. The marked drop in number of dinoflagellate species with relatively lower ratios of dinoflagellates to

sporomorphs, and increased abundance of vascular tissue and inertinite are considered to be palynological indicators of marine regression (Habib and Miller 1989; van Pelt 1990). Wall et al. (1977) showed that a species diversity index for Recent cysts tended to increase in offshore assemblages for transects in the Atlantic Ocean. May (1980) observed that relatively low to moderate species diversity and dominance by one or two species are characteristic features of assemblages from nearshore gulf (estuarine) sediments. Normal marine shelf assemblages tend to have moderate to high numbers of species with low dominance by one taxon. Edwards, Goodman and Witmer (1984) and Edwards (1989) concluded that a high species diversity indicated a marine shelf, whereas fewer species with a single dominant species in an assemblage represented coastal estuarine or brackish water environments.

Species diversity as a useful environmental criterion has been also applied to the C-T event. Tocher and Jarvis (1987) noted that the generally increasing trend in species diversity and numerical abundance upsection could be an indication of deepening water conditions. This cyst distribution pattern was also associated with an increase in the planktonic foraminifera (Hart and Bailey 1979; Hart 1982), a change in the inoceramid fauna (Jarvis and Woodrooff 1984), and a lithological change from the nodular chalks with hardgrounds into the flinty chalks (Tocher and Jarvis 1987). It was thought likely that all of the above features were related to

the major early-Turonian transgressive phase which has been recorded world-wide by numerous authors.

Relative Abundance of Species

The relative abundance of species or cyst morphotypes (groups) is the most widely used measurement to describe dinoflagellate assemblage composition, and to determine paleoenvironmental trends by changes in species associations through time. The technique has been used by palynologists studying Recent and fossil assemblages. Among numerous studies, either certain cyst morphotypes or certain cyst associations or ratio of specific morphotypes has been thought to be indicative of depositional environments.

Different dinocyst species or cyst morphotypes are relatively abundant or dominant in different environments (Wall et al., 1977). Vozzhennikova (1965) suggested that certain cyst morphotypes could be used to indicate depositional environments, with 'thick-walled' forms more abundant in 'neritic' environments (probably inner to middle neritic) and 'thin-walled process-bearing' forms more characteristic of 'open marine' conditions (probably outer neritic to upper slope). Scull et al. (1966) recognized distinctive associations of cysts as potential indicators of transgressive and regressive facies changes, based on the size and complexity of processes of certain skelochorate forms. The cysts with short, stout and barbed processes

(*Cleistosphaeridium*) appear to be restricted to near-shore brackish conditions, whereas the long, forked variety (*Polystephanephorus*) is indicative of a shallow marine environment. Scott (1982) suggested that smooth cysts are characteristic of nearshore, high-energy, turbid environments, whereas spiny cysts indicate deeper, low-energy, cleaner water. However, intermediate morphotypes are common and there is a complete continuum of assemblage types. This morphotype fractionation is probably purely hydrodynamic in origin (hydrodynamic sorting) (Scull et al., 1966; Davey 1971; Aurisano 1980, 1989). It has been recognized, even though not agreed by everyone, that the possession of long, complex processes (ornament morphology) as entangling devices indeed has a beneficial effect on the entrainment of the cyst in microconvection currents, thereby delaying the final sedimentation and promoting greater dispersal (Williams 1977; Tappan 1980; Sarjeant et al., 1987).

Downie et al. (1971) used relative abundance to define four cyst associations in the Lower Eocene of southern England. The associations usually occurred in particular facies, and therefore they were interpreted as being characteristic of particular paleoenvironments. The *Spiniferites* and *Areoligera* associations are interpreted to indicate 'open marine' conditions (probably outer neritic/upper slope), the *Micrhystridium* association inner neritic, and the *Wetzeliella* association lagoonal or brackish

water environments. Harland (1973) compared the relative abundance of two major groups of cysts by using the ratio of the number of gonyaulacacean cyst species to the number of peridiniacean cyst species, which he called the gonyaulacacean ratio. This ratio tended to be higher in more open marine environments in the Upper Cretaceous Bearpaw Formation of Alberta. These results and the findings of Downie and others (1971) tend to be supported each other, since *Spiniferites* is a gonyaulacacean cyst, whereas *Wetzeliella* is a peridiniacean type. The nearshore maxima of peridinoid dinocysts probably reflects a combination of their more euryhaline character and also their enrichment in areas of upwelling (Davey 1971; Wall et al., 1977; Bujak 1984). The results obtained from Recent offshore sediments off southwest Africa by Davey and Rogers (1975) indicated that *Spiniferites ramosus* predominated in sediments beneath regions of upwelled cold, nutrient-rich water. The study of Campanian-Maestrichtian dinoflagellates in southeastern Nigeria suggests that the assemblages with abundant *Spiniferites*-type cysts correspond to an inner-neritic environment, whereas the assemblages with relatively higher concentration of peridinoid cysts represent nearshore or transitional environments with slightly reduced salinity (Edet and Nyong 1993). Leckie et al. (1990) reported that the dominance of peridinoid *Chichaouadinium vestitum* in a low diversity assemblage indicates a restricted, marginal marine environment with slightly subnormal salinity in a shoreline

setting.

Detailed studies of the Cenomanian-Turonian dinoflagellate cyst associations in northern Europe have revealed that the cyst assemblages could be directly related to lithofacies and palynofacies (Marshall and Batten 1988). The *Cyclonephelium*/*Eurydinium* association only occurred in marls which yielded an abundance of amorphous organic matter. The *Spiniferites* association was most typical of the palynologically productive light grey marls and chalks but it also occurred in some dark marls which contained small to moderate amounts of amorphous detritus (Marshall and Batten 1988). The marls and chalks with *Spiniferites* in abundance suggested deposition in well-oxygenated, open marine water at a time when the oxygen minimum zone was restricted to the outer shelf. The replacement by the *Cyclonephelium*/*Eurydinium* association was considered to reflect a more highly stressed environment that resulted from a further extension of anoxic conditions into the euphotic zone possibly causing species of these genera to encyst. This may have been brought about by the development of relatively isolated basins which, with reduced circulation, would rapidly become anoxic, or by a substantial rise in the oxygen minimum zone across the shelf (Marshall and Batten 1988). The results of Eshet and others (1992) from the K-T boundary section of Israel tended to support this conclusion.

Absolute Abundance of Dinocysts

The absolute abundance of dinoflagellate cysts can be measured as the number of dinocysts per unit mass of sediment. It is also referred as the concentration of dinocysts in sediments. The concentration of dinocysts in marine sediments is controlled by a number of complex and interacting factors which include first the environmental and biological controls which affect the primary biogeographic distribution of the living species in the planktonic phase and its encystment in the benthic stage (production of cysts) and secondly, the factors which determine the behavior of both viable and non-viable cysts hydrodynamically as small organic particles in transportation and sedimentation, and the depositional environments for preservation (preservation of cysts). This technique has seldom been used for studying pre-Quaternary dinoflagellates, because it is difficult to distinguish the effects of the production factors from those of the preservation factors.

Factors influencing the distribution of dinoflagellates species in the plankton are numerous. These factors include autecological characteristics such as temperature and salinity tolerance ranges, nutrient requirement, motility and flotation properties, life-cycle features, as well as competitive characteristics and distance from the shoreline (Wall 1971). In addition, some environmental factors (those which may have a selective effect upon phytoplankton species composition in

a given water mass) are also listed: these include effects of horizontal transport and migration, bathymetric conditions and physiographic barriers (Wall 1971). The fact remains however, that the precise individual autecological preferences of the majority of modern dinoflagellate species are still poorly known or documented. Also, there is presently insufficient evidence to list all the factors which could induce cyst formation. It seems fairly clear that encystment of dinoflagellates is an adaptation to the unstable hydrographic conditions (Wall et al., 1977). Lowering nitrogen level is considered to be a factor inducing encystment (von Stosch 1973; Pfiester 1975, 1976, 1977; Walker and Steidinger 1979). Observation by Wall and Dale (1968) suggests that for a given species, encystment often begins around the peak of exponential growth and maximum abundance of its motile stages in the plankton, which has depleted nitrogen level significantly.

In general, the higher the motile dinoflagellate concentration, the larger the number of cysts produced. As with other microplankton, motile dinoflagellates are more-or-less related to primary productivity, which in turn is related to the availability of the nutrients (Davey 1971; Davey and Rogers 1975; Aksu et al., 1989; Hillaire-Marcel et al., 1989). It has been suggested that dinoflagellates are most abundant during the periods and within regions of upwelling (e.g., Dodson and Thomas 1977; Schnak and Elbrachter 1981). The study

by Melia(1984) indicated that the upwelling zone off the coast of West Africa was well defined by a pronounced north-south linear zone of dinoflagellate concentration which is parallel to shoreline. However, the evidence is still insufficient to conclude that maximum dinoflagellate proliferation corresponds directly to cyst abundance. The dinoflagellates which produce preservable cysts comprise only a small proportion of the total dinoflagellate population and hence it is not surprising that cyst abundance in the sediments is not directly related to the dinoflagellate abundance in the overlying waters (Davey and Rogers 1975). It remains a possibility that even in areas of low productivity, more dinoflagellate encystment will occur if a proper environmental condition exists (e.g., dramatical and frequent changes in temperature, or nitrogen levels, etc.).

Dinoflagellate cysts of the same general size and same approximate chemical composition as spores and pollen grains might be distributed quite differently from spores and pollen grains in the marine sediments (Traverse 1988). The abundance of spores and pollen grains in marine sediments is largely controlled by the distance away from the shoreline (decreasing sharply as moving offshore)(e.g., Bordovskiy 1965a, 1965b; Farley 1987), whereas dinocyst concentration in sediments is primarily controlled by the rate and density of cyst production by a parental population which itself is subject to numerous variables including seasonal ones, and influences

introduced subsequently such as dissemination, transportation, sedimentation and preservation. The sedimentation of individual cysts is dependent on several environmental variables such as currents, turbulence and the behavior of cysts as sedimentary particles in the hydrographic regime such as size, shape, specific gravity and condition (with or without protoplasts, living or dead) (e.g., Brideaux 1971a; Williams 1971a, 1971b; Wilson 1971; Wall and Dale 1973, 1974; Harker and Sarjeant 1975; Anderson et al., 1982; Harland 1983; Goodman 1979, 1987; Williams and Bujak 1985; Taylor 1987; Ford 1988).

Meroplanktonic dinoflagellate cysts, being very small organic particles, possess very low setting velocities and thus will settle only in the absence of turbulence. In the natural environment it may be expected that maximum abundances of dinocysts are found where there is high cyst production and where dinocysts can deposit without dispersal or winnowing of the sediment. Traverse and Ginsburg (1966, 1967) noted that the distribution of dinoflagellate cysts in the Bahamas generally followed that of other small organic particles (pine pollen, microforaminifera shell linings) and that they were concentrated in areas of non-turbulence characterized by lime muds. Depth is also important for dinocyst concentration in marine sediments because the proportion of dinocyst surviving sedimentation decreases with increasing residence time in the water column. Under oxygenated conditions, most dinocysts may

be destroyed when water depth is great enough. Recent studies for zooplankton fecal pellets reveal that most zooplankton (copepod) fecal pellets are produced by animals feeding on natural phytoplanktons. Smaller particulate matter ingested by zooplankton can be deposited and better preserved because copepod fecal pellets are relatively large and sink rapidly, and because the pellets are compacted in the gut of the animal and are protected by a mucilaginous or chitinous peritrophic membrane, which helps preserve the contents from biological, geochemical, or physical disruption and dissolution (Porter and Robbins 1981). The modern zooplankton fecal pellets are typically composed of algal remains, including a considerable part of motile dinoflagellates (Porter and Robbins 1981). The origin of amorphous debris is considered to be largely fragments of fecal pellets when surrounding peritrophic membrane is broken or dissolved (Porter and Robbins 1981). There is an apparent, albeit inconsistent, relationship between the richness of dinocysts and the development of amorphous debris (Habib and Miller 1989).

As has been summarized by many observers (e.g., Traverse 1988), the sedimentary situation into which palynomorphs are deposited has very great control over the likelihood of preservation: acid environments preserve better than alkaline, reducing environments better than oxidizing, and quiet sedimentary situations better than energetic ones. The exclusion of oxygen is particularly important. Since the

residence time of particulate organic matter in the bioturbated surface sediment layer is much longer than that in the water column, the most significant loss of organic matter by oxidative degradation should therefore be in the benthic transition layer rather than in the water column. There is general agreement that large quantities of organic matter can be preserved only in anoxic conditions that limit bacterial degradation by all but the anaerobic species and prevent auto-oxidative breakdown (e.g., Demaison and Moore 1980). Three basic models have been raised to reconstruct the environmental conditions under which organic matter accumulates. (1) Anoxic bottom conditions resulted from stagnation of a water column is a necessary prerequisite for the formation of organic-rich deposits. This belief is best exemplified by studies on the Black Sea sediments. Under such conditions, dissolved oxygen is used up in the decomposition of organic matter and is not readily replenished from surface layers. The much slower degradation of both marine and terrestrially derived organic matter due to the action of reducing bacteria leads to better preservation of sedimentary organic material; (2) The raised levels of organic matter in anoxic basins may represent faster rates of sedimentation rather than slower rates of decomposition. Large quantities of organic debris supplied at a sufficiently rapid burial rate may actually have caused anoxic conditions in the immediately overlying waters, no matter whether the original bottom water is anoxic or oxic

(Habib 1982); (3) In addition to restricted basins and marginal marine environments where sedimentation rates are high, anoxic conditions are also formed where a greatly increased productivity overwhelms the oxygen content of the water column, with the excess, unoxidized organic matter being incorporated into the sediment. The enhanced vertical circulation (particularly in some regions of coastal upwelling) quickly recycles nutrients into the surface productive zones. Mid-water oxygen-minimum zones are the most likely result of this scenario (Brass et al., 1982).

2. Depositional Organic Facies

The term palynofacies was introduced by Combaz (1964) and Batten (1982a, 1982b) to describe the total assemblage of microscopic organic matter present in a rock after maceration (in HCl and HF). It is not precisely synonymous with total kerogen content, generally defined as that part of organic matter which is insoluble in ordinary organic solvents (Durand 1980). The term depositional organic facies was proposed by Habib (1979, 1982, 1989). There is no fundamental difference between the two definitions, but depositional organic facies better represents the nature of total organic residues in thermally immature sediments. Three major types of organic matter found in marine sediments are palynomorphs, phytoclasts (various tissue and carbonized material), and amorphous particles. Generally, amorphous debris and oxidized particles

(inertinite) comprise the bulk of total organic matter concentration in typical marine sediments. Palynomorphs are considered to be a minor component in such an association. It is estimated that the preservation of organic matter under anoxic conditions results in assemblages overwhelmingly dominated by amorphous debris (70-90%) in all areas except those immediately adjacent to fluvial influx and/or loci of redeposition (Tyson 1985). The concentration of phytoclasts changes predictably in an offshore (proximal-distal) direction. With increasing distance from fluvial sources, prolonged reworking, exposure and oxidation, the land plant materials tend to become smaller, more equigranular, darker in color and highly oxidized, eventually becoming what is termed 'residual organic matter' (Tissot et al., 1980; Habib 1982). In the offshore areas, the allochthonous phytoclast fraction is masked by the much larger reservoir of marine carbon which becomes available under anoxic conditions. It appears that the oceanographic and hydrographic changes associated with bottom water oxygen depletion also frequently lead to a diminished supply of allochthonous materials to the basin (Tyson 1984, 1985). From the preceding discussion, it can be seen that the basic classification of organic matter generally reveals little about the paleoenvironments of marine organic-rich sediments except very general and relative estimates of proximity to fluvial sources, the degree of redeposition, and bottom water oxygenation. More refined paleoenvironmental

information can be derived only from the detailed examination of the palynomorph component of the organic residue.

Palynologically defined organic facies are classified on the basis of the dominant constituents in the organic residues (Fig. 3). The name of the facies usually follows the predominant component (Habib and Miller 1989). The vast majority of samples collected from continental margins and shelves (nearshore and offshore) contain organic residues with both terrestrial and marine origin. Terrestrially derived organic matter include pollen grains and spores (sporomorphs), vascular and other cellular tissues, fresh water algae, and oxidized or carbonized landplant materials (inertinite). Marine components include amorphous debris, dinoflagellate cysts and acritarchs.

(1) Amorphous Debris Facies

This facies is defined by the abundant well-preserved amorphous debris and paucity of land plant materials. It is considered to form in marine conditions because most of amorphous debris are the remains of fecal pellets produced by zooplankton digesting motile dinoflagellates and other microplankton. Dinoflagellate cysts are generally numerous and well-preserved, although predominant species vary from nearshore marine to farther offshore shallow shelf. Sporomorph species usually form a small percentage (<25%) of the total palynomorph assemblages (Habib and Miller 1989).

ORGANIC FACIES		DEFINITION	PALEOENVIRONMENTAL INTERPRETATION
Amorphous Debris Facies		-abundant well-preserved amorphous debris -diverse dinoflagellate assemblages -paucity of landplant particles	shallow shelf away from terrigenous detritus
Inertinite Facies	Coastal	-high concentration of Inertinites -poorly-preserved landplant materials -few dinoflagellate species(coastal) or no dinoflagellates(non-marine)	coastal environments: estuarine, tidal flat, lagoon interdistributary bay
	Non-Marine	-abundant oxidized amorphous debris	subaerial environments (excluding coal, peat, etc.)
Landplant Detritus Facies		-abundant well-preserved landplant vascular and cellular tissues -relatively abundant Inertinites -few dinoflagellate species with various numerical abundances	delta front, proximal prodelta, prodeltaic excursion onto shallow shelf

Fig. 3 Depositional organic facies and their environmental significance. After Habib and Miller 1989.

(2) Inertinite Facies

The inertinite facies is defined by the high concentration of inertinite. High content of inertinite may not be due to absolute increase but mainly to the absence of other organic particles (Habib and Miller 1989). Inertinite particles can be found in any organic facies, but only in the inertinite facies are vascular tissue and amorphous debris poorly represented and preserved, suggesting early diagenetic oxidation and/or recycling. The inertinite facies is thought to be related to nonmarine or estuarine or interdistributary bay coastal environments (Habib and Miller 1989).

(3) Landplant Detritus Facies

This facies is defined by the abundance of well-preserved vascular and other cellular tissue derived from land plants. Inertinite may be common (partly due to recycling), and amorphous debris is relatively little. Dinoflagellate diversity is usually low to moderate, but the abundance of specimens varies. The dinocyst assemblage can be represented by fewer species with larger numbers of individuals. This facies occurs in coastal areas close to a source of land vegetation and in nearshore shallow marine shelf environments (Habib and Miller 1989).

Cretaceous Western Interior Seaway

Geologic Setting

The Cretaceous seaway of the western interior North America was a broad, north-south elongated, west-east asymmetrical basin formed between late Jurassic to Cretaceous time in response to increased mid-ocean ridge spreading and high convergence and subduction rates along the western margin of the North American Plate. During the time of maximum transgression, this seaway extended 4800 km in a north-south direction with connections to the Boreal Ocean(present-day Arctic Canada) and the Tethys Sea (Gulf of Mexico). West to east the seaway had a maximum width of 1600 km, extending from Utah and Arizona to Iowa and Missouri (Fig. 4).

This tectonically complex cratonic basin, which developed to the east of the Cordilleran Thrust Belt, consisted of four tectono-sedimentologic zones (Kauffman 1977) (Fig. 5):

- (1) a rapidly subsiding western foreland basin probably produced by crustal loading and containing thick coarse clastics (Jordan 1981);
- (2) a relatively broad axial basin with a discontinuous forebulge zone located to its west, separating the axial basin from the foreland basin;
- (3) a tectonic hinge zone between the axial basin and eastern stable platform, characterized by moderate

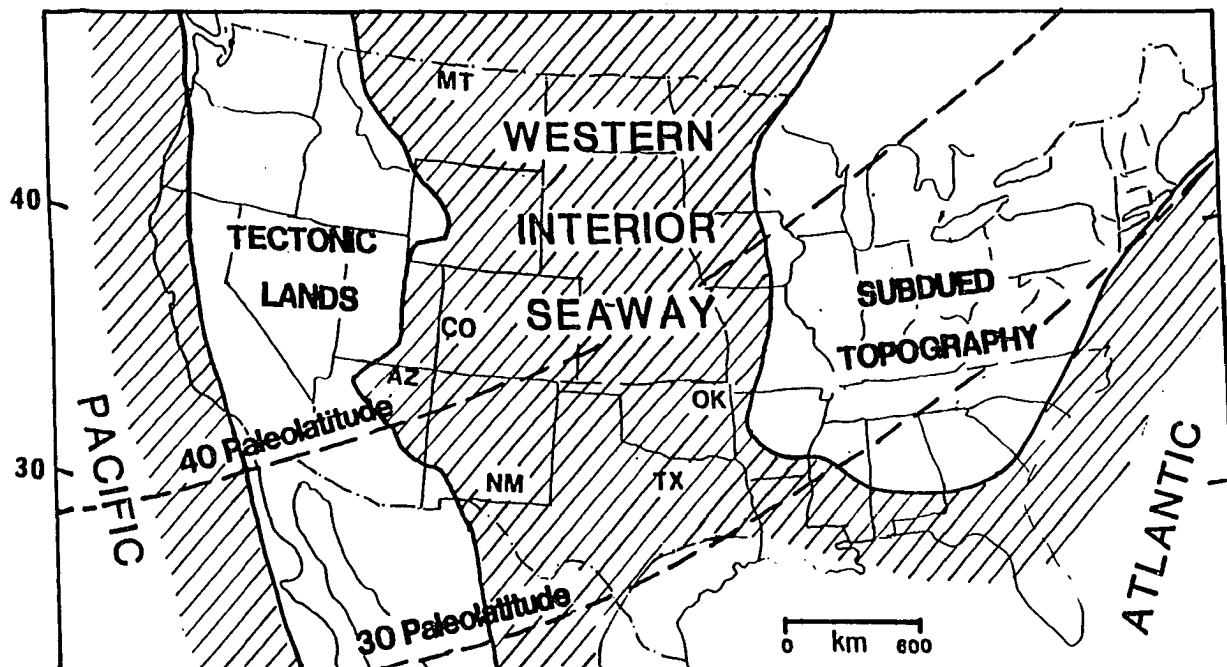


Fig. 4 Generalized paleogeographic map of the United States during the late Cenomanian-early Turonian time. After Hattin 1986.

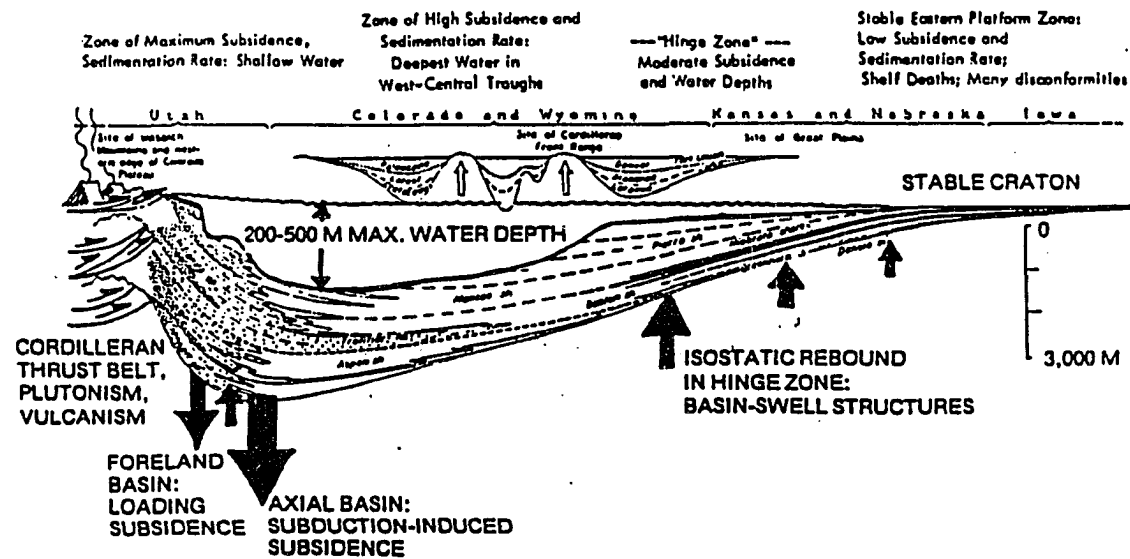


Fig. 5 Generalized structural and stratigraphic cross-section through the Western Interior Cretaceous Basin, showing major structural divisions of the basin. After Kauffman 1984b.

subsidence, episodic block uplifting or isostatic flexure;

(4) a broad eastern stable platform zone.

Episodic subsidence and block uplifting in the basin is considered to be related to periods of thrusting, plutonism, and volcanism in the Cordillera area, which are, in turn, apparently linked to accelerated seafloor spreading, convergence, and subduction along the western margin of North America (Kauffman 1984).

Second- through fourth-, and even fifth-order eustatic cycles are recorded from the latest Albian to middle Maestrichtian. However, five third-order eustatic fluctuations, which coincide with global third-order eustatic sealevel rises defined by Vail et al. (1977), are considered to be the primary control of sedimentation throughout the cratonal western interior basin (Kauffman 1985). The Cenomanian-Turonian marine transgression records the highest sea level rise (Greenhorn Cycle) during the late Albian to middle Maestrichtian time interval (Hancock and Kauffman 1979).

Oceanographic and Climatic Setting

During globally warm, equable, climatic episodes, such as persisted during Cretaceous high sea level periods, the atmospheric and oceanic circulation patterns may have been quite different from that of today. Warm saline bottom water

(WSBW), which is suggested to have formed in low-latitude epicontinental seas with a negative water balance (evaporation > precipitation + inflow), is considered to have been the major bottom water source instead of the modern cold polar bottom water (Brass et al., 1982). Since rates of WSBW formation are related to low-latitude shelf areas, which are determined by sea level, the vertical circulation in the ocean during times of peak transgression would be largely driven by increased WSBW formation (Brass et al., 1982). During times of decreased WSBW formation because of decreased subtropical shelf area or evaporation, oceanic circulation may have been sluggish and/or driven by the sinking of cold high-latitude waters (Brass et al., 1982; Saltzman and Barron 1982). Increased circulation, possibly as a result of increased WSBW formation at maximum transgression, may have intensified the oxygen-minimum zone, leading to the development of poorly oxygenated benthic conditions (Arthur et al., 1987).

Oceanic circulation in the western interior seaway of North America is somewhat different from that of open marine regions. Periodic variations in rainfall and runoff may have had a pronounced effect on water mass characteristics in the partially restricted seaway during maximum transgression. A low-salinity cap would have formed and isolated the bottom water mass from gaseous exchange with the atmosphere, resulting in anoxic bottom conditions (Arthur et al., 1987).

Precipitation patterns during the mid-Cretaceous have

been studied by several authors (e.g. Barron and Washington 1982). Unusually high global precipitation rates, particularly in the northern hemisphere mid-latitude continents, probably due to Cretaceous warmth and higher atmospheric CO₂ levels, are supported by the wide distribution of coal deposits, bauxites and kaolins (Fig. 6) (Barron and Washington 1982; Parrish et al., 1982; Parrish and Curtis 1982; Beeson 1984; Barron et al., 1985).

The cyclicity of the seaway stratigraphy is considered to be a reflection of Milankovitch-driven climatic alterations between wet and dry periods (Kauffman 1977; Pratt 1984, 1985; Arthur et al., 1985; Barron et al., 1985; Elder 1985, 1987b). In wet periods, high freshwater influx caused salinity stratification in the water column and stagnation in the bottom water, corresponding to clastic-rich deposits with high organic carbon content. In dry periods, vertical mixing of the water column occurred, corresponding to carbonate-rich deposits with low organic carbon content.

Lithofacies

Patterns of sedimentation in the western interior seaway were closely linked to both tectonic and tectonoeustatic controls (Kauffman 1977). At peak flooding (Cenomanian-Turonian), all major marine facies were developed in the basin. Elder (1987b) described 28 sections across the basin, most of which were located in western and central parts of the

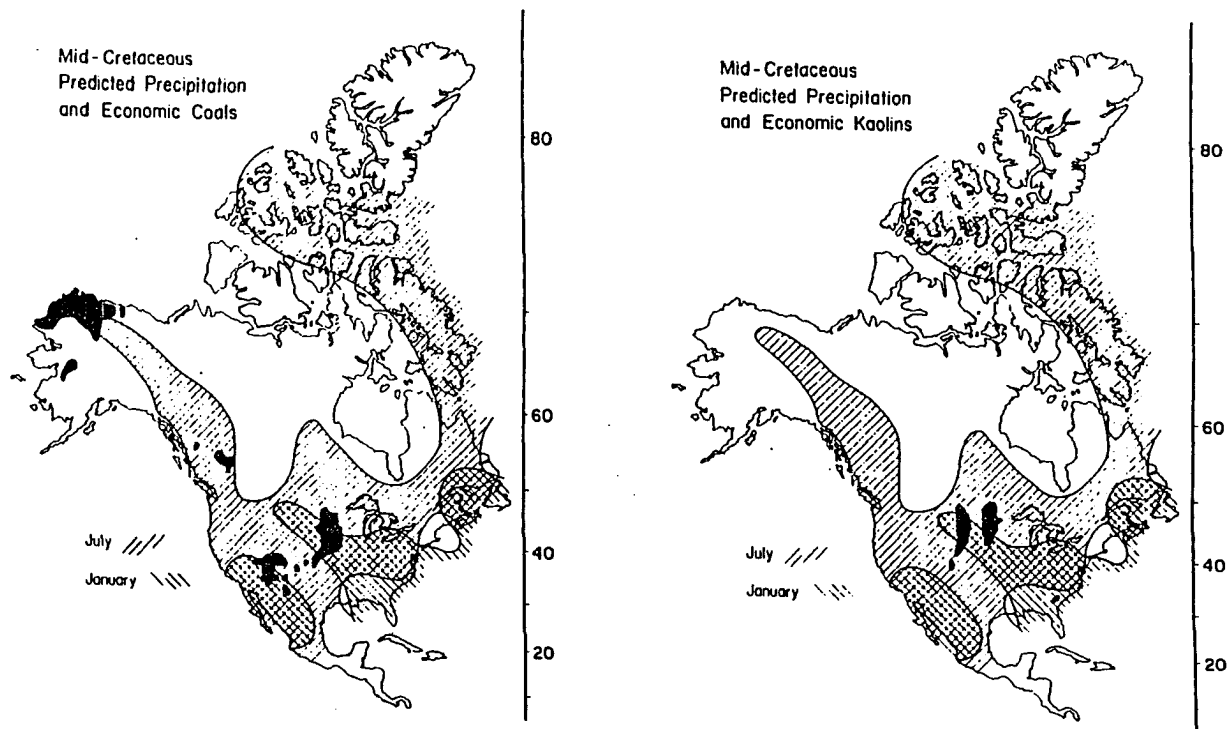


Fig. 6 Map of the distribution of mid-Cretaceous economic coal and kaolins and model-predicted high precipitation regions for North America. After Tourtelot 1983; Beeson 1984; Parrish and Barron, 1986.

seaway. He divided the strata into three major lithofacies suites, based on lithofacies associations and overall carbonate content. The carbonate content generally increases in offshore direction and through eustatic rise and transgression (Fig. 7 and Table 1). Each suite has its unique characteristics (Fig. 8). (1) Clay-dominated suite, which consists of non-calcareous to calcareous shales and claystones with some limestone concretions. This region represents nearshore localities (proximal shelf equivalent), where rock accumulation rates were either high(>6 cm/ka) in the western part or relatively low(2-3 cm/ka) in the eastern part; (2) Transitional suite, which is primarily composed of calcareous shales, marly shales and marlstones. This region is characterized by the continuous lithological changes from calcareous shales with limestone concretions in the late Cenomanian *Sciponoceras* zone, to marlstones and calcareous shales in the latest Cenomanian *Neocardioceras* zone, finally to admixture of calcareous shales, marly shales, marlstones and limestones in the early Turonian *Watinoceras* zone. This reflects intermediate offshore settings and sedimentation rates (3-6 cm/ka); (3) Carbonate-dominated suite, which consists of relatively thick limestones, marlstones and marly shales. This lithofacies suite is developed in central areas of the basin, where the sedimentation rates may be below 3 cm/ka. The boundaries of these suites vary somewhat in different periods due to an overall upsection increase in

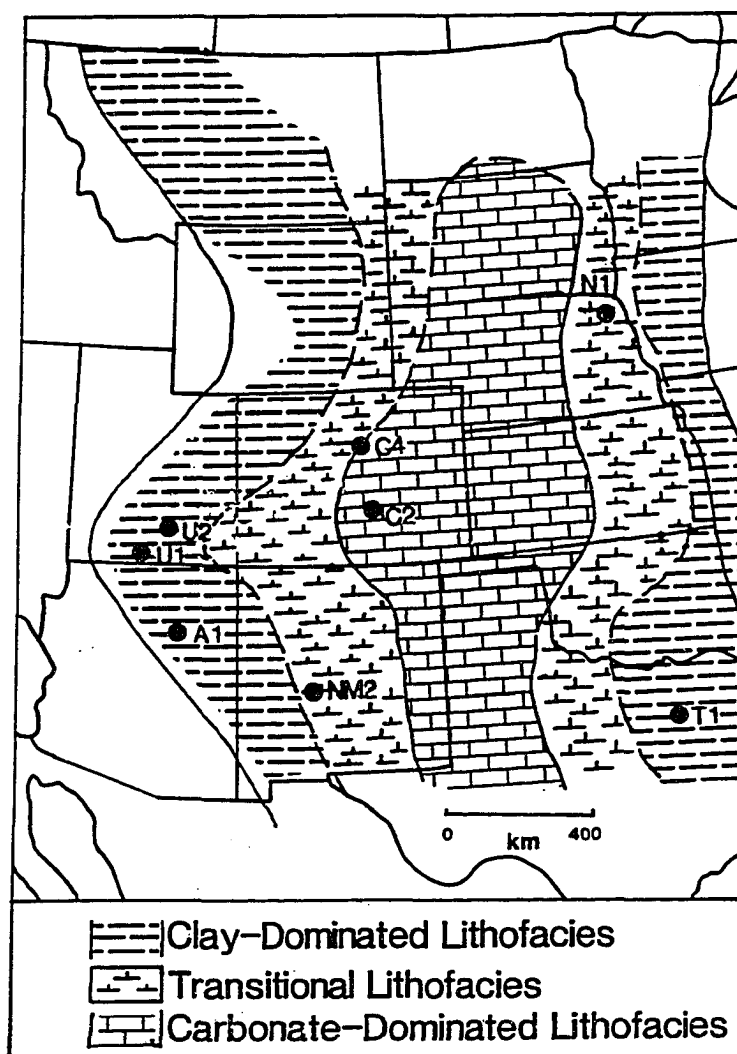


Fig. 7 Map of the Western Interior of the United States showing section locations and Cenomanian-Turonian shoreline and lithofacies distribution. Labeled section names mentioned in this study are A1=Blue Point, Arizona; C2=Pueblo, Colorado; C4=Boulder, Colorado; N1=Ponca, Nebraska; NM2=Carthage, New Mexico; T1=Dallas, Texas; U1=Wahweap Wash, Utah; U2=Egg Nog, Utah. After Elder 1987b.

SECTION		U1	A1	NM2	C2	N1	T1
CARBONATE CONTENT (%)	Watinoceras Zone	50	33	47	60	58	32
	Neocardioceras Zone	21	19	46	58	57	1
	Sciponoceras Zone	15	8	28	49	52	2
ORGANIC CARBON CONTENT (%)	Watinoceras Zone	0.84	0.95	0.50	2.38	1.80	2.87
	Neocardioceras Zone	1.20	0.96	0.55	2.46	4.30	0.95
	Sciponoceras Zone	0.70	0.46	0.45	0.70	5.40	0.93

Table 1 Average carbonate and organic carbon percentage in calcareous and marly shales at sections along the west-east transect showing in Figure 7. Limestone data excluded. Data from Elder 1987b.

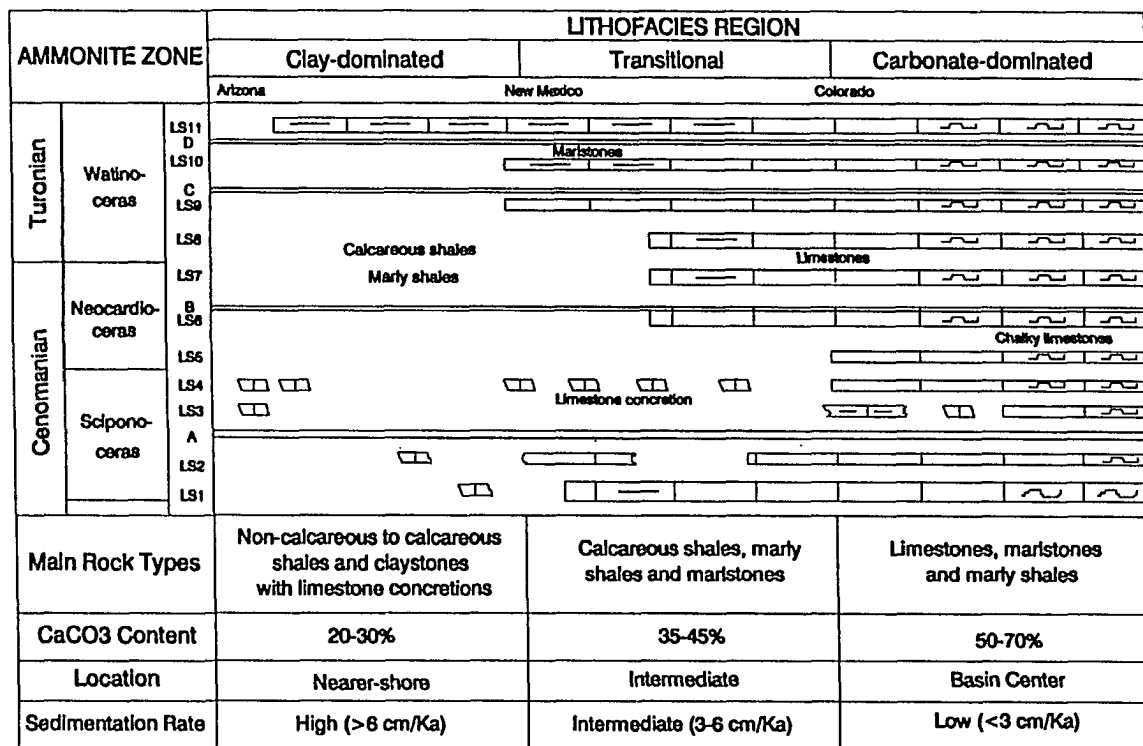


Fig. 8 Cross section along the transect from Arizona to Colorado showing the characteristics of three lithofacies suites. Modified from Elder (1987b).

carbonate content reflecting progressive transgression and deepening of the basin (Elder 1987b). The sedimentation rates change significantly across the western part of the basin, suggesting the more influence of tectonic activity and basinal subsidence (Elder 1987b).

Detailed stratigraphic analysis of alternating beds of burrowed limestones and laminated, organic-rich marlstones in the Bridge Creek Member of the Greenhorn Formation and its stratigraphic equivalents from Arizona and other areas (Fig. 9) have led to the suggestions that bedding couplets reflect the Earth's axial precession cycle (mean period 21 kyr), obliquity cycle (mean period 42 kyr), and orbital eccentricity cycle (mean period 100 kyr) (Gilbert 1895; Hattin 1971, 1979, 1985, 1986; Kauffman 1977; Fischer 1980; Fischer et al., 1985; Pratt 1981, 1984; Elder 1985, 1987a, 1987b, 1989). The relatively large, extensive limestone beds have been considered to have developed in a cyclicity of approximately 100 kyr (orbital eccentricity cycle) (Elder 1985) or of 41 kyr (obliquity cycles) (Fischer et al. 1985; Elder 1993, pers. comm.). Marlstone beds occurring between these major limestone units may represent the 21 and 41 kyr precession and obliquity cycles, respectively (Elder 1985).

SECTION	STUDY INTERVAL	DAKOTA FORMATION	DAKOTA SANDSTONE	MANCOS SHALE	TUNNCK SHALE MEMBER, Mancos Shale	LOWER SHALE MEMBER, Mancos Shale	BRIDGE CREEK LIMESTONE BEDS	GREENHORN LIMESTONE	HARTLAND SHALE MEMBER	BRIDGE CREEK MEMBER	Ponca Nebraska (N1)	Dallas Texas (T1)

Fig. 9 Stratigraphic nomenclature for strata in the study interval and area. After Elder 1987b.

Biostratigraphy

1. Macrofossils

The Cenomanian and Turonian Stages were originally recognized by d'Orbigny (1850) in the type areas of Sarthe and Touraine, respectively. The Stratotype and parastratotype sections of these stages include deposits formed during a series of successive transgressions. The C-T boundary was basically defined by ammonites and bivalves. Earlier studies placed the stage boundary between the *Sciponoceras gracile* Zone and *Watinoceras coloradoense* Zone in the U.S. Western Interior (Kennedy and Hancock 1977; Cobban and Scott 1977). Subsequent work led to, after the recognition of a new *Neocardioceras juddii* Zone, an elevation of the boundary from the top of *Sciponoceras gracile* to the top of the overlying *Neocardioceras juddii* or to the base of the *Pseudaspidoceras flexuosum* Zone (equivalent in time to the *Watinoceras coloradoense* Zone) (Hancock and Kennedy 1980; Wright and Kennedy 1981; Cobban and Hancock 1983; Cobban 1984, 1985; Kennedy and Cobban 1991) (Fig. 10).

2. Foraminiferas

Foraminiferal zonation of the C-T boundary interval is well established but does not offer high resolution. The boundary is located within the *Whiteinella archaeocretacea* Zone (Fig. 10), the base and top of which are defined by the

last occurrence of *Rotalipora cushmani* and the first occurrence of *Praeglobotruncana helvetica*, respectively (Eicher 1977; Robaszynski and Caron 1979; Wonders 1980; Leckie 1985).

3. Calcareous Nannofossils

There have been numerous published nannofossil biostratigraphic studies for the Upper Cretaceous (e.g., Thierstein 1976; Roth 1978; Watkins 1985, 1986, 1989; Bralower 1988). The C-T boundary was defined within the *Parhabdolothus asper* Zone. This zone was split into two subzones by the last occurrence of *Microstaurus chiastius* and the first occurrence of *Eiffellithus eximius* (Fig. 10). There is little agreement as to the exact nannofossil definition of the C-T boundary in different areas.

4. Dinoflagellates

A dinoflagellate zonation for the Cenomanian-Turonian interval was published more than two decades ago (e.g., Clarke and Verdier 1967). The stage boundary in offshore southeastern Canada is located between the *Kiokansium polypes* Interval Zone (Cenomanian) and the *Surculosphaeridium longifurcatum* Interval Zone (Turonian) (Fig. 10) (Williams 1975). The bottom of the *Kiokansium polypes* zone is defined by the last occurrence of *Chichaouadinium vestitum* (Bridueax) Bujak and Davies. This coincides with the result of Sverdløve and Habib (1974) that

C. vestitum is restricted to Albian sediments. *Dinopterygium cladoides* Deflandre and *Kleithriasphaeridium readei* (Davey and Williams) Davey and Verdier have their first occurrences within the *Kiokansium polypes* zone. The *Surculosphaeridium longifurcatum* zone is characterized by the first occurrence of *Senoniasphaera protrusa* Clarke and Verdier, and the last occurrence of *Litosphaeridium siphoniphorum* (Cookson and Eisenack) Davey and Williams and *Epelidosphaeridia spinosa* (Cookson and Hughes) Davey. Age control for such zones is based on planktonic foraminifera (Williams 1975, 1977).

The Cenomanian-Turonian dinocyst zonation in the English section is described by Clarke and Verdier (1967) as the *Litosphaeridium siphoniphorum* Concurrent Range Zone (Cenomanian) and the *Endoscrinium companulum* Assemblage Zone (early Turonian). The former is from the first occurrence of *Palaeohystrichophora infusorioides* Deflandre to the last occurrence of *L. siphoniphorum*. It is split into three subzones, namely *Maghrebinia perforata*, *Epelidosphaeridia spinosa* and *Cleistosphaeridium huguoniotii*. In Australia, the Cenomanian-Turonian dinocyst zone is the *Ascodinium parvum* Zone, which consists of the *Diconodinium dispersum* subzone (late Cenomanian) and the *Fromea fragilis* subzone (latest Cenomanian-early Turonian) (Norvick 1976).

Previous dinoflagellate studies of the critical Cenomanian-Turonian interval in the western interior of North America include those of Singh (1964, 1971, 1983; Brideaux

1971a) for Alberta, Canada, Jones (1976) for Black Mesa, Arizona, Pannella (1966) for the Denver Basin, Upshaw (1963, 1964), Griggs (1970), Burgess (1971), Sulkoske (1975), Nichols and Jacobson (1982, 1982b), Jacobson and Nichols (1982), and Nichols et al. (1982) for Wyoming. No previous works have been published on Cenomanian-Turonian strata in New Mexico, Nebraska, Texas and Utah.

MATERIALS AND METHODS

One hundred thirty samples from eight sections in the Western Interior basin were collected for palynological study (Fig. 11). They represent the sediments from three different lithofacies regions. The sedimentology, stratigraphy and paleontology of these sections has previously been investigated in detail, especially for those in the western part of the basin (Kauffman 1977; Elder 1985, 1987a, 1987b, 1989, 1990; Kirkland 1983, 1990, 1991).

Field work was carried out in June 1989. Samples from the Pueblo and Boulder Sections, Colorado were collected by the author. Others were provided by Dr. William Elder. Most of the sections were sampled in great detail, with an average of approximately one to two samples per meter. In the critical intervals, such as the C-T boundary and the *Sciponoceras-Neocardioceras* zonal boundary, two to three samples were collected per meter. Sample numbers are taken from Elder (1987b) except for those collected in the Boulder section (C4). Most samples consist of green to dark green, laminated and slightly bioturbated marly shales, calcareous shales and calcareous claystones.

All samples were processed by the standard palynologic method in the Palynology Laboratory at Queen College. Their

	A1	U1	U2	NM2	C4	C2	N1	T1
Watinoceras Zone	5	5	4	6	33 ¹ (12)	6	3	3
Neocardioceras Zone	6	6	3	5	14 ² (8)	5	3	2
Sciponoceras Zone	5	4	0	7		3	4	11
TOTAL	16	15	7	18	47³ (20)	14	10	16
1 may include some from the overlying biozone 2 the Sciponoceras zone and the Neocardioceras zone haven't been differentiated 3 47 —nmuber of sample collected; (20)—number of sample analysed								

Fig. 11 Sample analysed from eight sections for the study of dinoflagellate cysts and organic facies.

mineral components were largely or entirely removed by digesting in 5% HCl and in concentrated 48% HF. Organic residues were mounted for light microscopy. Commonly, four slides were made for each sample: The first was prepared from unfiltered residues for the analysis of depositional organic facies. The filtration through a 20 μ m sieve was taken to concentrate palynomorphs. Two slides were prepared from particles greater than 20 μ m and the fourth one was made from the fraction smaller than 20 μ m. All dinocysts in the two +20 μ m slides were counted. The number of specimens is usually 200-1000 in the calcareous shales and less than 100 in the claystones. For some samples from the C2, C4 and N1 sections (all with relatively higher carbonate content), there are too few specimens even though additional 10-20 slides were counted. Relative abundance of each species was described according to the following scheme: very rare (1-2 specimens counted); rare (3-5 specimens); few (6-15 specimens); common (16-25 specimens); and abundant (greater than 25 specimens).

The preservation of dinocysts is highly variable and shows a trend of fewer specimens in the more calcareous rocks. As discussed in the previous chapter, there are many factors affecting cyst production, transportation and deposition. Therefore, the concentration of dinocysts in sediments is rarely used to infer the change of depositional environments. This study will only discuss the role of carbonate content in the concentration of dinocysts in the C-T sediments. Without

question, different lithotypes will also affect species diversity. However, many studies have showed that species diversity is more related to the change of environments mostly resulted from the expansion and contract of the continental shelf during marine transgression and regression (e.g., Partridge 1976; Habib and Miller 1989). The recent study shows that a very similar change in species diversity from the different assemblages in the different lithofacies regions is considered to be mostly the result of marine transgression and regression.

Table 2 lists the 151 taxa of dinoflagellates and acritarchs (including species, subspecies and unnamed cyst-forms) that were identified in this study. Approximately 85% of species were observed in the clay-dominated A1 section, and to a lesser content, in the transitional NM2 section. As much as 26% of the total taxa listed in Table 2 have not been formally described.

In this study, the ratio of dinoflagellates to sporomorphs is not measured due mainly to a very low concentration of the sporomorphs in most samples. Sporomorphs, most of which are bissacate pollen grains and fern spores, are generally only 5 percent or even less in the assemblages. Other landplant materials, including vascular and other cellular tissues, are also relatively few and poorly preserved (fragmented and oxidized).

The analysis of depositional organic facies is based on

Table 2 List of dinoflagellate and acritarch species in the investigated sections.

Achomosphaera ramulifera (Deflandre 1937) Evitt 1963
Achomosphaera neptuni (Eisenack 1958) Davey and Williams 1966
Achomosphaera sagena Davey and Williams 1966
Achomosphaera spp.
Apteodinium granulatum Eisenack 1958
Apteodinium sp.
Ascodinium acrophorum Cookson and Eisenack 1960
Ascodinium parvum (Cookson and Eisenack 1958) Cookson and Eisenack 1960
Avellodinium sp.
Batiacasphaera euteiches (Davey 1969) Davey 1979
Batiacasphaera reticulata (Davey 1969) Davey 1979
Baltisphaeridium spp.
Boukidinium psilatum Singh 1983
Chytroeisphaeridium baetica Riegel 1974
Chytroeisphaeridium minor (Cookson and Hughes 1964) Morgan 1980
Cleistosphaeridium aciculare Davey 1969
Cleistosphaeridium sp. cf. *C. armatum* (Deflandre 1937) Davey 1969
Cleistosphaeridium clavulum (Davey 1969) Below 1982
Cleistosphaeridium huguoniotii (Valensi 1955) Davey 1969
Cleistosphaeridium multispinosum (Singh 1964) Brideaux 1971
Cleistosphaeridium parvum Davey 1969
Cleistosphaeridium spp.
Cordosphaeridium sp.
Coronifera minus (Yu and Zhang 1980) Mao and Norris 1988
Coronifera oceanica Cookson and Eisenack 1958
Coronifera striolata (Deflandre 1937) Stover and Evitt 1978
Coronifera sp.
Cyclonephelium compactum Deflandre and Cookson 1955
Cyclonephelium distinctum Deflandre and Cookson 1955
Cyclonephelium eisenacki Davey 1969
Cyclonephelium hystrix (Eisenack 1958) Davey 1978
Cyclonephelium membraniphorum Cookson and Eisenack 1962
Cyclonephelium paucispinum Davey 1969
Cyclonephelium vannophorum Davey 1969
Cyclonephelium sp.
Cymatiosphaera radiata (O. Wetzel 1933) Sarjeant 1985
Deflandrea sp.
Dichardogonyaulacysta sp.
Diconodinium sp.
Dinocyst sp. A
Dinocyst sp. B
Dinocyst sp. C
Dinocyst sp. D
Dinocyst sp. E
Dinocyst sp. F
Dinocyst sp. G
Dinocyst sp. H
Dinogymnium sp.
Diphyes cretaceum Yu and Zhang 1980
Diphyes minor Yu and Zhang 1980
Diphyes sp.
Epelidosphaeridia spinosa (Cookson and Hughes 1964) Davey 1969
Exochosphaeridium bifidum (Clarke and Verdier 1967) Clarke et al. 1968
Exochosphaeridium phragmites Davey et al. 1966
Florentinia cooksoniae (Singh 1971) Duxbury 1980
Florentinia deanei (Davey and Williams 1966) Davey and Verdier 1973
Florentinia ferox (Deflandre 1937) Duxbury 1980
Florentinia mantellii (Davey and Williams 1966) Davey and Verdier 1973
Florentinia rediculata (Davey and Williams 1966) Davey and Verdier 1973
Florentinia sp.
Fromea amphora Cookson and Eisenack 1958
Fusiformacysta sp.
Geiselodinium sp.
Ginginnodinium evittii Singh 1983
Gonyaulacysta cassidata (Eisenack and Cookson 1960) Sarjeant 1966
Gonyaulacysta sp. cf. *G. cretacea* (Neale and Sarjeant 1962) Sarjeant 1969
Gonyaulacysta helicoidea (Eisenack and Cookson 1960) Sarjeant 1966
Gonyaulacysta spp.
Hapsocysta sp.
Heterosphaeridium conjunctum Cookson and Eisenack 1968
Heterosphaeridium sp.
Hystrichodinium pulchrum Deflandre 1935
Hystrichosphaeridium bowerbankii Davey and Williams 1966
Hystrichosphaeridium difficile Manum and Cookson 1964

Diphyes minor Yu and Zhang 1980
Diphyes sp.
Epelidosphaeridia spinosa(Cookson and Hughes 1964) Davey 1969
Exochosphaeridium bifidum(Clarke and Verdier 1967) Clarke et al. 1968
Exochosphaeridium phragmites Davey et al. 1966
Florentinia cooksoniae(Singh 1971) Duxbury 1980
Florentinia deanei(Davey and Williams 1966) Davey and Verdier 1973
Florentinia ferox(Deflandre 1937) Duxbury 1980
Florentinia mantellii(Davey and Williams 1966) Davey and Verdier 1973
Florentinia rediculata(Davey and Williams 1966) Davey and Verdier 1973
Florentinia sp.
Fromea amphora Cookson and Eisenack 1958
Fusiformacysta sp.
Geiselodinium sp.
Ginginnodinium evittii Singh 1983
Gonyaulacysta cassidata(Eisenack and Cookson 1960) Sarjeant 1966
Gonyaulacysta sp. cf. *G.cretacea*(Neale and Sarjeant 1962) Sarjeant 1969
Gonyaulacysta helicoidea(Eisenack and Cookson 1960) Sarjeant 1966
Gonyaulacysta spp.
Hapsocysta sp.
Heterosphaeridium conjunctum Cookson and Eisenack 1968
Heterosphaeridium sp.
Hystrichodinium pulchrum Deflandre 1935
Hystrichosphaeridium bowerbankii Davey and Williams 1966
Hystrichosphaeridium difficile Manum and Cookson 1964
Hystrichosphaeridium recurvatum(White 1842) Davey and Williams 1966
Hystrichosphaeridium stellatum Maier 1959
Hystrichosphaeridium tubiferum(Ehrenberg 1838) Deflandre 1937
Hystrichosphaeridium sp.
Isabelidinium magnum(Davey 1970) Stover and Evitt 1978
Isabelidinium sp.
Kallosphaeridium sp.
Kiokansium erectum(Manum and Cookson 1964) Below 1982
Kiokansium hesperum(Davey 1979) Stover and Williams 1987
Kiokansium polypes(Cookson and Eisenack 1962) Below 1982
Kiokansium williamsii Singh 1983
Kiokansium sp.
Litosphaeridium conispinum Davey and Verdier 1973
Litosphaeridium stiphoniphorum(Cookson and Eisenack 1958) Davey and Williams 1966
Luxadinium sp.
Micrhystridium fragile Deflandre 1947
Micrhystridium inconspicuum(Deflandre 1935) Deflandre 1937
Micrhystridium stellatum Deflandre 1945
Micrhystridium spp.
Muderongia perforata Alberti 1961
Odontochitina costata Alberti 1961
Odontochitina operculata(O.Wetzel 1933) Deflandre and Cookson 1955
Odontochitinaopsis sp.
Oligosphaeridium anthophorum(Cookson and Eisenack 1958) Davey 1969
Oligosphaeridium complex(White 1842) Davey and Williams 1966
Oligosphaeridium sp. cf. *O.dispare*(Tasch 1964) Davey and Williams 1969
Oligosphaeridium sp. cf. *O.pulchrrimum*(Deflandre and Cookson 1955) Davey and Williams 1966
Ovoidinium verrucosum(Cookson and Hughes 1964) Davey 1970
Palaeohystrichophora infusorioides Deflandre 1935
Palaeoperidinium sp. cf. *P.cretaceum* Pocock 1962.
Pareodinia sp.
Pervosphaeridium brevispinum(Norvick 1976) Below 1982
Pervosphaeridium cenomaniense(Norvick 1976) Below 1982
Pervosphaeridium pseudohystrichodinium(Deflandre 1937) Yun 1981
Pervosphaeridium sp. cf. *P.truncatum*(Davey 1969) Below 1982
Prolixosphaeridium sp. cf. *P.conulum* Davey 1969
Pseudoceratium eisenackii(Davey 1969) Bint 1986
Pterodinium cingulatum(O.Wetzel 1933) Below 1981
Pterodinium cingulatum granulatum(Clarke and Verdier 1967) Lentin and Williams 1981
Pterospermella australiensis(Deflandre and Cookson 1955) Eisenack et al. 1973
Pterospermella hartii(Sarjeant 1960) Eisenack et al. 1973
Pterospermella pastiei(Durand 1958) Eisenack et al. 1973
Pterospermella sp.
Scriniodinium campanula Gocht 1959
Senoniasphaera sp.
Sentusidinium eisenackii(Boltenhagen 1977) Lentin and Williams 1973
Sentusidinium spp.
Spiniferites ancoriferus Cookson and Eisenack 1974
Spiniferites cornutus laevimurus(Davey and Williams 1966) Lentin and Williams 1973
Spiniferites multibrevis(Davey and Williams 1966) Below 1982
Spiniferites ramosus brevifurcatus(Cookson and Eisenack 1974) Lentin and Williams 1977
Spiniferites ramosus gracilis(Davey and Williams 1966) Lentin and Williams 1973
Spiniferites ramosus granomembraneus(Davey and Williams 1966) Lentin and Williams 1973
Spiniferites ramosus ramosus(Ehrenberg 1838) Deflandre 1937
Spiniferites ramosus reticulatus(Davey and Williams 1966) Lentin and Williams 1973
Spiniferites sp. cf. *S.?spiciosus*(Deflandre 1937) Sarjeant 1970
Spiniferites sp. cf. *S.wetzelii*(Deflandre 1937) Sarjeant 1970
Spiniferites spp.
Subtilisphaera deformance(Davey and Verdier 1973) Stover and Evitt 1978
Subtilisphaera foliacea Eisenack and Cookson 1960

Isaberridium sp.
Kallosphaeridium sp.
Kiokansium erectum (Manum and Cookson 1964) Below 1982
Kiokansium hesperum (Davey 1979) Stover and Williams 1987
Kiokansium polypes (Cookson and Eisenack 1962) Below 1982
Kiokansium williamsii Singh 1983
Kiokansium sp.
Litosphaeridium conispinum Davey and Verdier 1973
Litosphaeridium siphoniphorum (Cookson and Eisenack 1958) Davey and Williams 1966
Luxadinium sp.
Micrhystridium fragile Deflandre 1947
Micrhystridium inconspicuum (Deflandre 1935) Deflandre 1937
Micrhystridium stellatum Deflandre 1945
Micrhystridium spp.
Muderongia perforata Alberti 1961
Odontochitina costata Alberti 1961
Odontochitina operculata (O. Wetzel 1933) Deflandre and Cookson 1955
Odontochitinopsis sp.
Oligosphaeridium anthophorum (Cookson and Eisenack 1958) Davey 1969
Oligosphaeridium complex (White 1842) Davey and Williams 1966
Oligosphaeridium sp. cf. *O. dispere* (Tasch 1964) Davey and Williams 1969
Oligosphaeridium sp. cf. *O. pulcherrimum* (Deflandre and Cookson 1955) Davey and Williams 1966
Ovoidinium verrucosum (Cookson and Hughes 1964) Davey 1970
Palaeohystrichophora infusorioides Deflandre 1935
Palaeoperidinium sp. cf. *P. cretaceum* Pocock 1962.
Pareodinia sp.
Pervosphaeridium brevispinum (Norvick 1976) Below 1982
Pervosphaeridium cenomaniense (Norvick 1976) Below 1982
Pervosphaeridium pseudohystrichodinium (Deflandre 1937) Yun 1981
Pervosphaeridium sp. cf. *P. truncatum* (Davey 1969) Below 1982
Prolixosphaeridium sp. cf. *P. conulum* Davey 1969
Pseudoceratium eisenackii (Davey 1969) Bint 1986
Pterodinium cingulatum (O. Wetzel 1933) Below 1981
Pterodinium cingulatum granulatum (Clarke and Verdier 1967) Lentin and Williams 1981
Pterospermella australiensis (Deflandre and Cookson 1955) Eisenack et al. 1973
Pterospermella hartii (Sarjeant 1960) Eisenack et al. 1973
Pterospermella pastielsi (Durand 1958) Eisenack et al. 1973
Pterospermella sp.
Scriniodinium companula Gocht 1959
Senoniasphaera sp.
Sentusidinium eisenackii (Boltenhagen 1977) Lentin and Williams 1973
Sentusidinium spp.
Spiniferites ancoriferus Cookson and Eisenack 1974
Spiniferites cornutus laevimurus (Davey and Williams 1966) Lentin and Williams 1973
Spiniferites multibrevis (Davey and Williams 1966) Below 1982
Spiniferites ramosus brevifurcatus (Cookson and Eisenack 1974) Lentin and Williams 1977
Spiniferites ramosus gracilis (Davey and Williams 1966) Lentin and Williams 1973
Spiniferites ramosus granomembraneus (Davey and Williams 1966) Lentin and Williams 1973
Spiniferites ramosus ramosus (Ehrenberg 1838) Deflandre 1937
Spiniferites ramosus reticulatus (Davey and Williams 1966) Lentin and Williams 1973
Spiniferites sp. cf. *S. ?spiciosus* (Deflandre 1937) Sarjeant 1970
Spiniferites sp. cf. *S. wetzeli* (Deflandre 1937) Sarjeant 1970
Spiniferites spp.
Subtilisphaera deformance (Davey and Verdier 1973) Stover and Evitt 1978
Subtilisphaera foliacea Eisenack and Cookson 1960
Subtilisphaera pirnaensis (Alberti 1959) Jain and Millepie 1973
Subtilisphaera pontis-mariae (Deflandre 1936) Lentin and Williams 1976
Subtilisphaera senegalensis Jain and Millepie 1973
Subtilisphaera trendellii (Cookson and Eisenack 1970) Lentin and Williams 1976
Subtilisphaera ?ventriosa (Alberti 1959) Jain and Millepie 1973
Subtilisphaera sp.
Surculosphaeridium longifurcatum (Firtion 1952) Davey et al 1966
Systematophora cretacea Davey 1979
Tanyosphaeridium salpinx Norvick 1976
Tanyosphaeridium variecalamum Davey and Williams 1966
Trichodinium sp. cf. *T. castaneum* (Deflandre 1935) Clarke and Verdier 1967
Trithyrodinium suspectum (Manum and Cookson 1964) Davey 1969
Veryhachium reductum (Deunff 1959) Downie and Sarjeant 1965
Veryhachium spp.
Xenascus ceratioides (Deflandre 1937) Lentin and Williams 1973
Xenascus blastema (Davey 1970) Stover and Helby 1987
Xiphoniphorum alatum Sarjeant 1966

examination of the unfiltered slide. This method was initially proposed by Habib (1979) in DSDP site study, and revised in Cretaceous-Tertiary Atlantic Coastal Plain sediments by Habib and Miller (1989). The definition and their paleoecological significance are illustrated in Figure 3.

RESULTS OF INVESTIGATION

Dinoflagellate stratigraphy across the C-T boundary is discussed within the context of lithofacies regions based on the presence of relatively consistent palynologic characteristics in each region. Characteristics include the composition of assemblage, species diversity and specimen abundance. Eight sections from three different lithofacies regions will be discussed lithologically (also see Appendix B) and then the dinoflagellate stratigraphy will be followed by an analysis of depositional organic facies and a general interpretation of environments.

1. Clay-dominated Lithofacies Region

Three sections were analyzed in the clay-dominated lithofacies region (Fig. 7). One lies in the Black Mesa Basin of northeastern Arizona (A1); another lies in the Kaiparowits Basin of southern Utah (U1); and another lies near Dallas, Texas (T1). The clay lithofacies region contains the most diverse and abundant macrofossils (Elder 1987b) and microfossils (including dinoflagellates) and, therefore, provides the most complete data on the study of the C-T boundary.

(i) Black Mesa, Arizona (A1)

The section in the Black Mesa Basin lies at or very near the base of the Mancos Shale (Fig. 9). It is considered to represent a nearshore setting about 100-150 km distant from the western shoreline of the seaway (Elder 1987b). Previous work has been published by Kirkland (1983, 1990, 1991) for stratigraphy and paleontology; Pratt and Threlkeld (1984) and Pratt (1985) for stable carbon and oxygen-isotope and organic-carbon studies; Jones (1976) for dinoflagellate biostratigraphy; Bralower (1988) for calcareous nannofossils; and Elder (1987a, 1987b, 1989, 1991) for a complete discussion of the biotic (mainly macrofaunal) pattern across the C-T boundary.

Lithology

Most of the sediments in the *Sciponoceras* zone are characterized by moderately-well laminated, olive gray, silty, calcareous claystones situated above a poorly sorted, silty, calcareous, medium-grained basal sandstone. The smaller amount of quartz silt higher in the section was considered to be the double effect of decreased offshore sediment bypass and winnowing, and increased clay supply as transgression slowed and sedimentation rates increased (Elder 1987b; Kirkland 1990). The stratigraphically higher *Neocardioceras* zone is characterized by a similar association of rocks. Silt content and induration increase at or just above the C-T boundary in

the basal *Watinoceras* zone, and continue to increase upsection. Strata of the *Watinoceras* zone (only lower and middle parts) are characterized by calcareous, moderately-well indurated, moderately-well laminated, olive gray to olive black, shales or claystones (Elder 1987b; Kirkland 1990). Both carbonate content and organic carbon content increase upsection (Table 1).

Sixteen samples were collected from a 13-meter-thick section, five within the *Sciponoceras* zone, six within the *Neocardioceras* zone, and five within the *Watinoceras* zone, respectively. Most samples are olive gray to dark gray, silty, calcareous shales and claystones.

Dinoflagellate Stratigraphy

I. *Sciponoceras* Zone

Samples from the *Sciponoceras* Zone (late Cenomanian) yielded dinoflagellate cyst assemblages consisting of several hundred specimens (Fig. 12). Eighty-eight species and subspecies were recorded. Approximately half of them range through the entire section or through most of it. Assemblages invariably contain typically mid-Cretaceous species of *Cyclonephelium*, *Odontochitina*, *Florentinia*, *Spiniferites* and *Oligosphaeridium*. *Spiniferites ramosus ramosus*, *Cyclonephelium distinctum*, *C. vannophorum* and *Florentinia mantellii* are the most abundant forms. Other common long-ranging taxa include

Baltosphaeridium sp. B
Cheliosphaeridium multispinum
Eocheliosphaeridium pinagrites
Cheliosphaeridium sp. A
Cheliosphaera polyseta cf. *C. armatum*
Cheliosphaera sp. B
Cymbalosphaera radiata
Michysphaeridium incraspinatum
Xenascus blastema
Dinozyx sp. D

Dinozyx sp. A
Cheliosphaeridium sp. B
Spiriferites ramosus gracilis
Dinozyx sp. E
Michysphaeridium stellatum
Fluviomeryx sp. A
Michysphaera sp. B
Coronella costulata
Michysphaeridium minor
Pseudodinium paucispinum
Pseudodinium cingulatum cingulatum
Cymbalosphaeridium membrum
Epiliosphaera cf. *E. spinosa*
Pseudocerasium elenaeckii
Spiriferites sp. B
Michysphaeridium sp. B
Xenascus caraboides
Eocheliosphaeridium bifidum
Spiriferites sp. A
Spiriferites sp. C
Pervosphaeridium pseudohysterochordulum
Amphosphaeridium sp.
Amphosphaeridium alphonisporum
Leptosphaeridium cf. *L. catenatum*
Trichosphaera cf. *T. catenatum*
Dinozyx sp. F

Hydrisphaeridium cf. *H. difficile*
Cheliosphaeridium eocissare
Cheliosphaeridium bealesi
Subtilisphaera foliacea
Modiolium hesperium
Pervosphaeridium brevispinum
Odonotrochopsis sp.
Dinozyx sp. G
Forerella sp.
Umbellula sp.
Galeosodulum sp.
Coronella minor
Serolisidium sp. A
Gonyaulosphaera heliocides
Gonyaulosphaera sp. B
Pervosphaeridium subop.
Pseudodinium sp. A
Pseudodinium cingulatum granulosum
Isabelidium sp. A
Oligosphaeridium sp.
Scrinodinium campanula

Leptosphaeridium cf. *T. varicosulatum*
Pseudosphaeridium cf. *P. conulum*
Subtilisphaera sp. A
Heliosphaeridium cospidum
Acetansulum sp.
Subtilisphaera sp.
Serolisphaera sp.
Gonyaulosphaera sp.
Michysphaeridium sp.
Cymbalosphaera sp.
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Exochosphaeridium phragmites, *Kiokansium polypes*, *Oligosphaeridium complex* and *Spiniferites multibrevis*. Among the relatively short-ranging biostratigraphically useful dinoflagellate species are *Florentinia cooksoniae*, *Muderongia pariata*, *Pervosphaeridium brevispinum*, *P.pseudohystrichodinium* and *Xenascus ceratioides*. These taxa have range bases or tops within the *Sciponoceras* zone. Thirty-eight species and subspecies are restricted to the *Sciponoceras* zone, and of these, 24 have been recorded from post-Cenomanian sediments in European sections (Clarke and Verdier 1967; Jarvis et al., 1989). Fifty taxa, including almost all long-ranging forms, disappear temporarily in the upper part of the zone, and reappear in the overlying *Neocardioceras* zone.

II. *Neocardioceras* Zone

The *Neocardioceras* Zone yielded 77 species and subspecies (Fig. 12). *S.ramosus ramosus*, *C.distinctum* and *C.vannophorum* remain abundant. Other common forms within this zone include *K.polypes*, *E.phragmites*, *Achomosphaera ramulifera*, *Pervosphaeridium pseudohystrichodinium* and *P.brevispinum*. Taxa which first appear in the *Neocardioceras* zone include *Senoniasphaera sp.*, *Scriniodinium campanula*, *Coronifera minor* and *Prolixosphaeridium conulum*. Twenty-eight species and subspecies are restricted in this zone, including *Florentinia spp.*, *Coronifera spp.*, *Kiokansium williamsii*, *Sentusidinium spp.* and *Cleistosphaeridium clavulum*. Nearly 40 forms

temporarily disappear in the upper part of the zone (immediately prior to or precisely at the C-T boundary), but reoccur in the overlying *Watinoceras* zone.

III. *Watinoceras* Zone

This zone (lower and middle parts) yielded 62 species and subspecies (Fig. 12). Of these, 12 range through the C-T boundary, and 37 reappear at the beginning of this zone after a disappearance near the boundary. Another 12 species, including several forms of *Subtilisphaera*, first appear in this zone. The dinoflagellate assemblages in the lower part of the *Watinoceras* Zone are characterized by *O.operculata*, *O.costata*, *E.phragmites*, *K.polypes* and *C.distinctum*. Species of two peridinoid genera, *Subtilisphaera* and *Palaeohystrichophora*, such as *S.deformance*, *S.foliacea*, *S.pirnaensis* and *P.infusorioides* which appear or reappear above the C-T boundary, become increasingly common upwards, and predominate in the upper part of the *Watinoceras* Zone. Stratigraphically useful species like *Epelidosphaeridia spinosa* and *S.campanula* last appear in this zone.

IV. Discussion

Dinoflagellate cyst assemblages from the Arizona section yielded 127 species and subspecies through the C-T boundary interval. Nearly 80-90 taxa are widely-distributed forms which have been previously recorded in many European sections. Among

them, *Oligosphaeridium* complex, *Odontochitina operculata* and *O.costata* were considered to be relatively cosmopolitan (open marine) forms (Tocher and Jarvis 1987; Leckie et al., 1990). Approximately 40 species range through most of the section. No particular form was noted to be dominant in the assemblages through the entire section. The relatively abundant forms are *C.distinctum*, *C.vannophorum*, *S.ramosus ramosus* and *E.phragmites*. Other common types are *Florentinia* spp. and *O.complex* in the *Sciponoceras* zone, *K.polypes*, *P.brevispinum* and *P.pseudohystrichophora* in the *Neocardioceras* zone, and *Subtilisphaera* spp., *P.infusorioides*, and *Odontochitina* spp. in the *Watinoceras* zone.

The dominance in an assemblage by one or several species is low. However, the dominance by a couple of selected cyst-based groups in the associations is relatively high. The assemblages are dominated by the *Cyclonephelium* group over most of the studied interval. The *Spiniferites* group is much less abundant than *Cyclonephelium* group except in the lower part of the *Sciponoceras* zone where there is almost no difference in numerical abundance between the two groups. In most samples, the relative abundances of these two groups seem to be inversely related, with one increasing as the other decreases. Compared with the *Spiniferites* and *Cyclonephelium* groups, the peridinioid group is much less abundant over most of the section. However, the number of peridinioid cysts increases significantly in the middle part of the *Watinoceras*

zone, along with a decrease in the abundance of both *Cyclonephelium* and *Spiniferites* groups.

Dinoflagellates have been used to date Cenomanian-Turonian strata (e.g., Jarvis et al., 1988a). According to the standard biohorizons compiled by Chen et al. (see Haq et al., 1988), the last occurrence of *Litosphaeridium siphoniphorum* and the first occurrence of *Chatangiella* spp. and *Hystriosphæridium difficile* identify the early Turonian. Other characteristic datums include the last occurrence of *Trithyrodinium suspectum* (late Cenomanian age), and the last occurrence of *Epelidosphaeria spinosa* and *Scriniodinium campanula*, and the first occurrence of *Senoniasphaera protrusa* suggesting an early Turonian age. In the Arizona section, the important datums are based on last-occurrence(LO) datums. The last occurrence of *T.suspectum* in the *Sciponoceras* zone suggests a late Cenomanian age. The last occurrence of *E.spinosa* and *S.campanula* at the base of the *Watinoceras* zone indicates an age not older than latest Cenomanian. *Hystriosphæridium difficile* and *Chatangiella* spp. were not recorded in this section.

Both dinoflagellate diversity and specimen abundance decrease significantly adjacent the late Cenomanian *Sciponoceras*-*Neocardioceras* biozonal boundary and adjacent the C-T boundary. These coincide with the pronounced decreases in diversity of both benthic and planktonic faunas which are considered to be the major steps of the Cenomanian-Turonian

mass extinction (Kauffman 1977, 1984a, 1984b; Elder 1987a, 1987b, 1989, 1991). Most dinoflagellate species disappear temporarily in these two intervals, but then reappear in the above strata. It is noteworthy that the early event (43 species with more than 500 specimens dropping to four species with less than ten individuals) in the late Cenomanian seems to be more extensive than the later one (34 species with more than 300 specimens to 22 species with less than 100 individuals) prior to or at the C-T boundary. The macrofaunal and foraminiferal data also suggested a greater diversity loss across the *Sciponoceras-Neocardioceras* zonal boundary than the C-T boundary (Elder 1987b). A clear recovery is found immediately following the drops. Both decreases and recoveries are thought to be closely related to the sea level change and paleoceanographical change.

Comparison of Dinoflagellate Assemblages

(1) Northwestern Black Mesa Basin

Jones (1976) described the dinoflagellate assemblages from four sections of the Mancos Shale in the northwestern Black Mesa Basin. The Upper Cenomanian-Middle Turonian interval yielded 46 species of dinoflagellate and 24 species of acritarchs. Three dinocyst zones were established (Jones 1976). The results show a similarity to those recorded in present study (southwestern Black Mesa Basin). The dominant

species in the present study (*S. ramosus ramosus*, *C. distinctum*, *K. polypes*, *O. operculata* and *P. infusorioides*) are also the most abundant types in the Jones' study. However, differences between the two areas are also apparent in: (1) more diverse dinoflagellate assemblages are recorded in the southwestern section than northwestern ones (more than 120 species and subspecies versus average fifty); (2) there is a well-balanced dinoflagellate association (proximate, chorate and cavate cysts) in the present study, whereas in previous data, cavate cysts *P. infusorioides* and *Deflandrea spp.* (including several recent cavate genera) are extremely abundant in most intervals, including many high chorate cyst (*Spiniferites-Achomosphaera* complex) assemblages; and (3) contrary to the present study, where there is an apparent drop in dinoflagellate diversity around the C-T boundary, the species richness in the NW Black Mesa Basin remains same or slightly increases across the boundary. The reason why Jones' data did not show apparent fluctuations in diversity might be due to his broader sampling intervals. His sampling intervals are large, e.g., one sample per six/seven meters.

(2) Northwest Europe

It appears that there is a fair measure of agreement between the general dinoflagellate zonation proposed for northwest Europe (Clarke and Verdier 1967; Foucher 1980, 1981, 1982; Tocher and Jarvis 1987; Jarvis et al., 1987, 1988a,

1988b) and the stratigraphy recorded in the Western Interior basin. However, data presented in previous palynological studies of Cenomanian-Turonian boundary sequences in the Chalk Sea of Northwest Europe, including the Isle of Wight (Clarke and Verdier 1967), Pas-de-Calais (Foucher 1982), Devon (Tocher and Jarvis 1987; Jarvis et al., 1987, 1988b), and Dover, England (Jarvis et al., 1988a) indicate that some important differences exist between the dinoflagellate assemblages recorded by this study and those in coeval sediments from other parts of the world.

Dinoflagellate assemblages in NW European sections are much less diverse than in the Arizona section, and have different species dominance. In the Dover sequence (mostly chalks and marls), no one assemblage contains more than 20 dinoflagellate species in a total of only 37. The associations in Devon consist of a total 50-60 species, but only few of them produce more than 30 species. The Cenomanian assemblages in Dover are dominated by *Oligosphaeridium* complex, *Odontochitina costata*, *O. operculata*, *Cyclonephelium distinctum* and *Batiacasphaera euteiches*, and the Turonian associations are characterized by *C. distinctum*, *O. complex*, *B. euteiches* and *Senoniasphaera rotundata* (Jarvis et al., 1988a). In other European sequences, *P. infusorioides* is the most abundant type throughout the C-T boundary interval. The species which are abundant in some intervals include *Spiniferites* spp., *O. complex*, *Cyclonephelium membraniphorum*, *C. distinctum*,

O.costata, *S.rotundata* and *Canningia colliveri* (Clarke and Verdier 1967; Foucher 1980; Tocher and Jarvis 1987; Jarvis et al., 1987, 1988b).

A significant decline in dinoflagellate diversity and abundance near the C-T boundary is also recorded in the European sections, which is coincident with increased $\delta^{13}\text{C}$ values and macrofossil extinctions. In the Dover sequence, the C-T boundary is devoid of dinocysts, and the disappearance of most species in the latest Cenomanian is accompanied by increased proportions of *Kallosphaeridium ringnesiorum* (Manum and Cookson) Tocher and Jarvis. This thick-walled proximate species suddenly disappears prior to the C-T boundary just after reaching a maxima (constituting more than 95% of the cysts). This coincides with the peak of $\delta^{13}\text{C}$ excursion, suggesting the most intense expansion of oxygen-minimum layer (Jarvis et al., 1988a). Kauffman and others called this species a disaster species (pers. comm. 1989) because of its extreme abundance while other species become extinct or disappear. It is interesting that this dinocyst event has not been recorded in the other European localities (Jarvis et al., 1988a) or in Arizona. Both species diversity and abundance recover in almost all European sections after the C-T boundary, but they never reach the level of the latest Cenomanian.

Depositional Organic Matter

The lower part (most of the *Sciponoceras* zone) of this section is characterized by abundant amorphous debris, diverse dinoflagellate assemblages and a lower concentration of land-plant materials (Fig. 13). It gradually changes into an admixture of amorphous debris and inertinite particles at the top of this zone, even though the latter is still a minor component in the assemblages. The percentage of amorphous debris increases at the beginning of the *Neocardioceras* zone along with a burst of dinoflagellate species after a relatively lower concentration at the base, and remains same throughout the zone. Pollen grains and spores are very few, even in the intervals with relatively higher land-plant materials comprised mostly of oxidized and fragmented vascular tissues. An increase of land-plant materials occurs prior to and at the C-T boundary although in much lower amounts than at the *Sciponoceras*-*Neocardioceras* zonal boundary. Most of the *Watinoceras* zone (studied) is characterized by relatively constant organic facies similar to that in the preceding *Neocardioceras* zone. Dinoflagellate diversity fluctuates from relatively high in the lower part to moderate in the upper part.

Environmental Interpretation

The section in the onshelf area of Black Mesa Basin is the best representative of dinocyst occurrence and organic

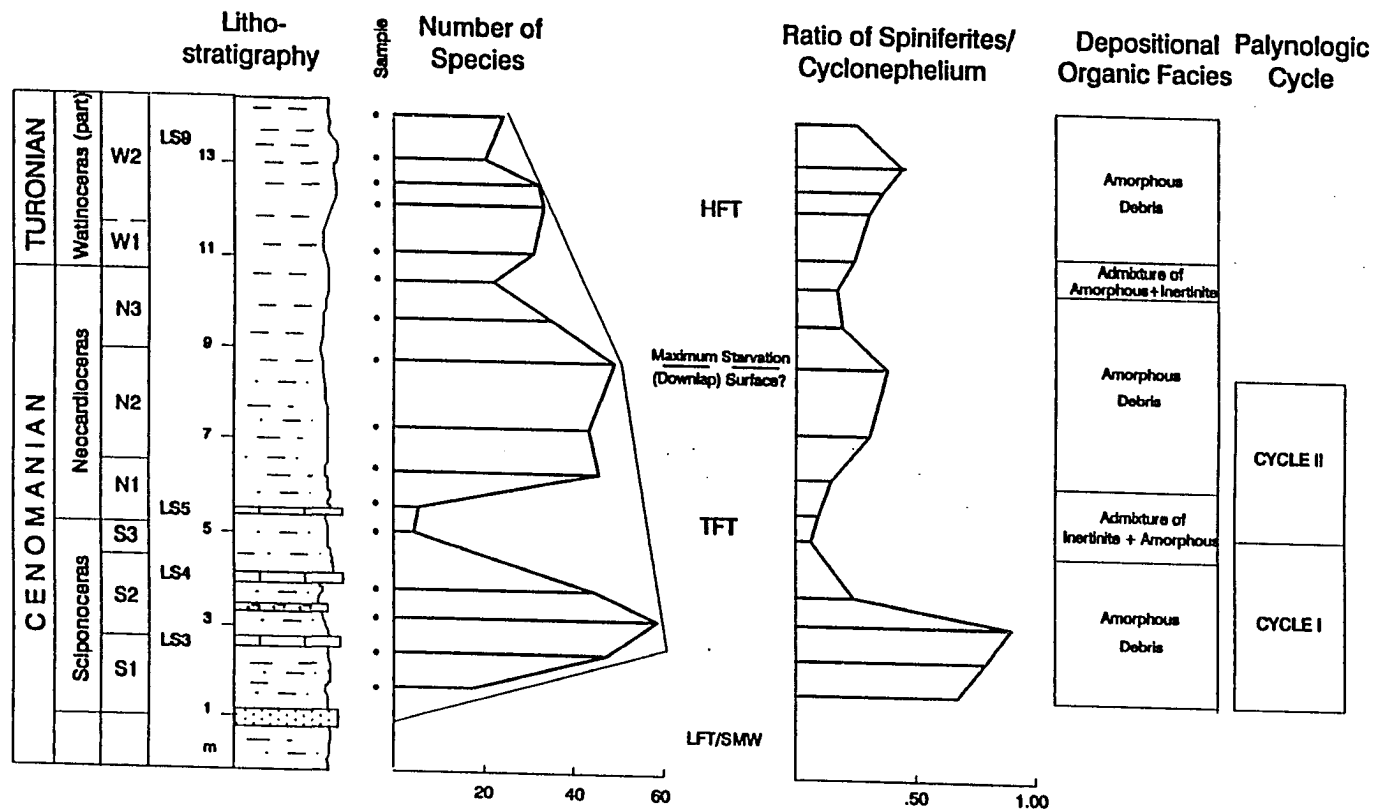


Fig. 13 Distribution of dinocyst species diversity, S/C ratio, and depositional organic facies in the Black Mesa, Arizona (A1).

facies. Figure 13 illustrates dinoflagellate diversity and the ratio of *Spiniferites* to *Cyclonephelium*. This is compared with the lithostratigraphy and depositional organic facies. The ratio of *Spiniferites/Cyclonephelium* (S/C ratio) is used for the first time as a criterion to indicate the change of depositional environments. The higher ratio represents offshore and the more open marine environments.

The distribution of dinoflagellates shows at least two major peaks in the sequence which correlates well with the sedimentological, macrofaunal and organic facies evidence. The first peak occurs within the *Sciponoceras* Zone where a less diverse dinoflagellate assemblage (17 species) at the bottom changes dramatically to highly diverse associations in the middle part (around 50 species). The increases in both species diversity and specimen abundance, their association with amorphous fecal debris, and the high S/C ratios suggest a sea level rise across the Black Mesa Basin. Habib et al. (1991, 1992, 1994) reported a similar trend in the Cretaceous-Tertiary (K-T) boundary interval in the Atlantic Coastal Plain, and interpreted it to be the result of a rapid expansion of the continental shelf during marine transgression. Sea level rise and rapid marine transgression in the Black Mesa Basin in early *Sciponoceras* time is also indicated by the transgressive lag sands at the base of the *Sciponoceras* zone, and the influx of diverse *Sciponoceras* faunas into Western Interior seaway from south (Kauffman 1984;

Elder 1987a, 1987b, 1991; Kirkland 1990, 1991). This lag sand in the nearshore area is considered to represent the transgressive surface (Elder 1987b). A dramatic decrease in dinoflagellate species and specimen abundance occurs at the top of the zone defined by the admixture of inertinite detritus and amorphous debris. That only four species with fewer than 30 individuals occurs in this interval, is contrasted by 43 species with more than 500 specimens in the preceding assemblages. The smaller number of dinoflagellate species, associated with both lower S/C ratio and the ratio of dinoflagellate to sporomorph species, and the occurrence of abundant inertinite and degraded land-plant materials is characteristic of coastal paleoenvironment which suggests a possible lowstand of sea level. Recent studies of modern and ancient estuarine sediments indicate that the dinoflagellate assemblages in estuarine environments consist only of a few species, and are usually overwhelmingly dominated by a single species (McMinn 1991; Leckie et al., 1990; Leckie and Singh 1991; Habib and Miller 1989; Habib et al., 1992, 1994). An abrupt negative $\delta^{18}\text{O}$ excursion at that time suggests a reduced salinity in the upper water column during stillstand and possibly short-term regression period, which is also coincident with the first major step of macrofossil extinctions and a loss of benthic foraminifera (Pratt 1985; Leckie 1985; Elder 1987a, 1987b).

The palynologic response to the second pulse(basal

Neocardioceras zone) of sea level rise and marine transgression is the significant increase in the number of dinoflagellate species, the recurrence of a higher S/C ratio and the abundance of amorphous debris in the lower part (Unit 21) of the zone. Multiple biostratigraphic datums are identified by the multiple occurrences of species immediately above poorly fossiliferous samples (Unit 18)(Li 1991). Such multiple occurrence datums contain both dinoflagellate species which range throughout, and species with the first or last appearances within the interval. This transgressive pulse at the base of the *Neocardioceras* zone is indicated also by a minor disconformity at the base of the *Neocardioceras* zone, condensation of strata with the concomitant shift of the center of deposition from the western to the eastern margin of the Black Mesa Basin, and the immigration of the *Neocardioceras* zone faunas (Elder 1987a, 1987b; Kirkland 1990, 1991). The other parts of the *Neocardioceras* zone are characterized by relatively diverse assemblages associated with the amorphous debris facies. Relatively higher organic content and a large negative $\delta^{18}\text{O}$ excursion suggest a continuous effect of fresh-water inputs at this time (Pratt 1985). The second apparent decrease in species diversity, associated with a decrease in S/C ratio, occurs prior to or just at the C-T boundary. With respect to the magnitude of decrease in species diversity and specimen abundance, the C-T boundary event is not as great as that at the *Sciponoceras*-

Neocardioceras zonal boundary. Less inertinites and degraded land plant materials, and more diverse dinoflagellate assemblages suggest that the change of depositional environment resulting from a drop in relative sea level occurs on a smaller scale. Many important dinoflagellate species either range across the boundary, or temporarily disappear prior to the boundary, and then reappear immediately in the above strata. Unlike macrofossils, dinocysts do not show any level of mass extinction near the C-T boundary. The study (Elder 1987a, 1987b) of lithofacies and macrofaunas indicated that decreased nearshore substrate stability due to relatively higher sedimentation rates, which was in response to increased rainfall and fluvial input, probably was an important factor to cause environmental deterioration in the *Neocardioceras* zone. Large parts of benthic and nekto-benthic taxa, especially bivalves and ammonites, were susceptible to the stresses responsible for the C-T boundary mass extinctions (Elder 1987a, 1987b). Planktonic diversity, on the other hand, did not vary as significantly as it did in benthic forms within the phase of environmental changes (Olesen 1991).

A rapid recovery in species diversity of dinoflagellates immediately above the C-T boundary reflects the third transgressive pulse. This is indicated by a sharp increase in carbonate content, increased faunal abundance, sediment bypass and sediment condensation in more offshore areas of the Black Mesa basin where the entire record of the uppermost

Neocardioceras zone is locally absent (Elder 1987a, 1987b; Kirkland 1990, 1991). Moderately fossiliferous dinoflagellate assemblages and abundant amorphous debris suggest that the sediments of the *Watinoceras* zone (lower and middle parts) were basically deposited in an open marine environment. However, the continued decreasing trend in species diversity and an increasing dominance of peridinioid cysts, such as *Palaeohystrichophora* and *Subtilisphaera*, may indicate the movement towards nearshore condition probably resulted from increased influence of shoreline progradation.

In conclusion, the sequence of depositional organic facies and dinoflagellate diversity in the Arizona section (A1) suggests cycles of marine transgression and regression responding to change of sea level. The model that is used is that first proposed by Habib et al. (1992) for the basal Danian of Alabama. In sequence stratigraphy, the Alabama study showed that the fewest dinoflagellate species occur in the Lowstand Facies Tract and in the uppermost part of the Highstand Facies Tract. The number of species is highest in the lower part of the transgressive interval, directly above the lowstand interval. The number of species remains high in the transgressive interval but decreases systematically into the highstand facies, and becomes lowest at the top. Based on the sequence stratigraphy published by Baum and Vail (1988), this pattern of dinoflagellate diversity is considered to represent low diversity in relatively nearshore environments

(LFT), and then dramatic increase in diversity with sea level rise across a broad shelf environment (TFT) followed by a reduction in diversity as the result of shoreline progradation (HFT).

In the Arizona section, the late Cenomanian interval (basal *Sciponoceras* zone) is characterized by the admixture of inertinite and amorphous debris, deposited in nearshore environments at the beginning of the transgression during the period of a relatively low sea level. This was followed by the amorphous debris facies, which is characterized by abundant well-preserved amorphous organic matter and only very little land-plant detritus, but diverse and numerous dinocysts. It was deposited in a rapid transgression. The subsequent regressive phase is represented by a decrease in diversity and an increase in inertinite and other terrigenous plant particles. A second palynological episode of transgression occurred again in the late Cenomanian (*Neocardioceras* zone). A possible regression adjacent the C-T boundary led to an apparent decrease in species diversity, although it was not as pronounced as the preceding late Cenomanian one. The relatively constant and rich amorphous debris facies, and diverse dinoflagellate assemblages throughout the *Watinoceras* zone suggest normal offshore marine conditions. This interpretation is supported by the abrupt increase in foraminiferal abundance (Leckie 1985; Leckie et al., 1991) in this interval.

The maximum diversity of dinoflagellates occurs near the base of the transgressive intervals, and not in the peak transgressions. For example, the largest number of species do not occur in the early Turonian where higher content of carbonates and maximum influx of subtropical faunas imply deeper water conditions than earlier in the study interval (Elder and Kirkland 1985). Rather, maximum dinoflagellate diversity occurs at the beginning of transgression at the base of *Sciponoceras* zone (Fig. 13).

Figure 14 illustrates the sequence of diversity in the Al section correlated against the biostratigraphy and lithostratigraphy of the section and the sequence stratigraphy of the late Cenomanian-early Turonian published by Haq et al. (1988). The overall trend in number of dinoflagellate species shows that it is largest in the late Cenomanian early transgressive phase (*Sciponoceras* zone) and then decreases gradually through the early Turonian *Watinoceras* zone. This pattern corresponds with the finding shown for the basal Danian in Alabama (Habib et al., 1992), and indicates an overall transgression during the period of sea level rise shown by Haq et al. (1988). The smaller intervals of transgression and regression in the Al section, pronounced in the *Sciponoceras* zone, may represent flooding intervals corresponding to parasequence sets. It is noted that the generally decreasing trend in species richness from the beginning of transgressive facies tract to the highstand

Cenomanian-Turonian Sequence Stratigraphy and Eustatic-cycle (Haq et al., 1988)

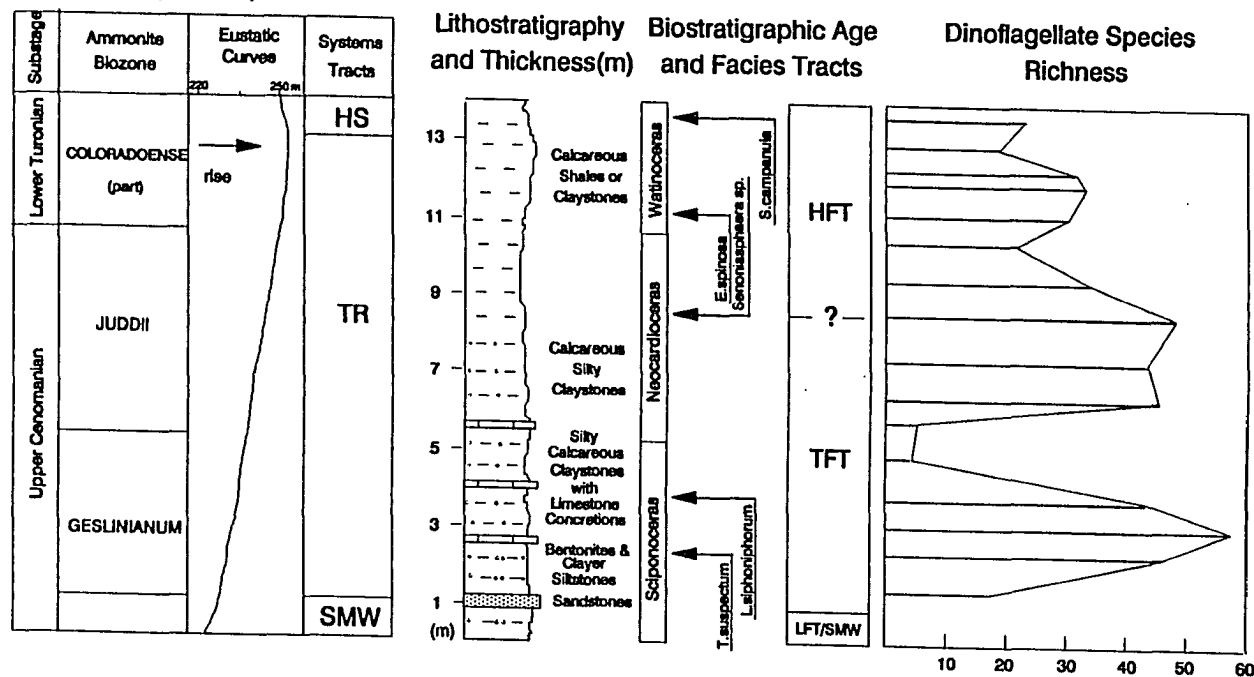


Fig. 14 Comparison of lithostratigraphy, palynology and inferred dinoflagellate sequence stratigraphy in Black Mesa, Arizona with standard sequence stratigraphy and eustatic-cycle after Haq et al. 1988.

facies tract exhibits an apparent inflection in the upper part of the *Neocardioceras* zone from which the decrease is getting more apparent. This inflection point in the curve of species diversity may reflect the boundary between the transgressive and highstand facies tracts. A similar phenomenon was also noted in the Braggs K-T section of Alabama (Habib et al., 1992). It is interesting that the inflection in the decreasing line in both sections occurs in the condensed intervals. It may be related to increased shoreline progradation.

(ii) Wahweap Wash, Utah (U1)

The Wahweap Wash section lies in the Tropic Shale of the Kaiparowits Basin (Fig. 9), and is probably located 100-150 km from the western margin of the seaway during deposition of strata in the study interval (Elder 1987b). Geologic, stratigraphic, and macrofaunal studies were conducted by Peterson (1969), Peterson and Kirk (1977), Elder (1987b) and Gustason (1989).

Lithology

The basal *Sciponoceras* zone is characterized by silty to sandy claystones and limestone concretions. The rest of the *Sciponoceras* zone consists mainly of poorly- to moderately-indurated, olive gray, silty, calcareous claystones interbedded with two limestone concretion beds (LS3 and LS4) in the lower and middle parts of the zone, respectively. The silt

content decreases upsection within this zone. The *Neocardioceras* zone is characterized by a similar combination of rocks as in the preceding zone with a silty shale bed at the base. Silt content decreases again upsection which suggested an increased winnowing or sediment bypass near the base (Elder 1987b). The middle part of the *Neocardioceras* zone is marked by a prominent calcareous marlstone bed (Unit 27). Above this resistant unit, silt content increases abruptly associated with a decrease in organic carbon content and a change in faunal composition near the C-T boundary. Elder (1987b) considered these phenomena to be the results of increased sedimentation rates and increased fluvial detrital supply. Higher sedimentation rates were indicated by the greater thickness of strata between marker-beds B and C (including the upper part of the *Neocardioceras* zone and the lower and middle parts of the *Watinoceras* zone) than the stratigraphically equivalent part in the Black Mesa (Elder 1987b). The strata of the *Watinoceras* zone studied consist of well indurated, olive black, very slightly silty, calcareous to marly, shales.

The organic carbon content is less than 1.0% in the *Sciponoceras* zone, 1.0-1.5% in the *Neocardioceras* zone, and less than 1.0% in the *Watinoceras* zone. Carbonate content is relatively low (15 to 21%) through the *Sciponoceras* and *Neocardioceras* zones, but increases abruptly at the base of the *Watinoceras* zone (50%) (Elder 1987b).

Fifteen samples were analysed, including four within the *Sciponoceras* zone(5-meter-thick), six within the *Neocardioceras* zone(7-meter-thick), and five within the *Watinoceras* zone(11-meter-thick).

Dinoflagellate Stratigraphy

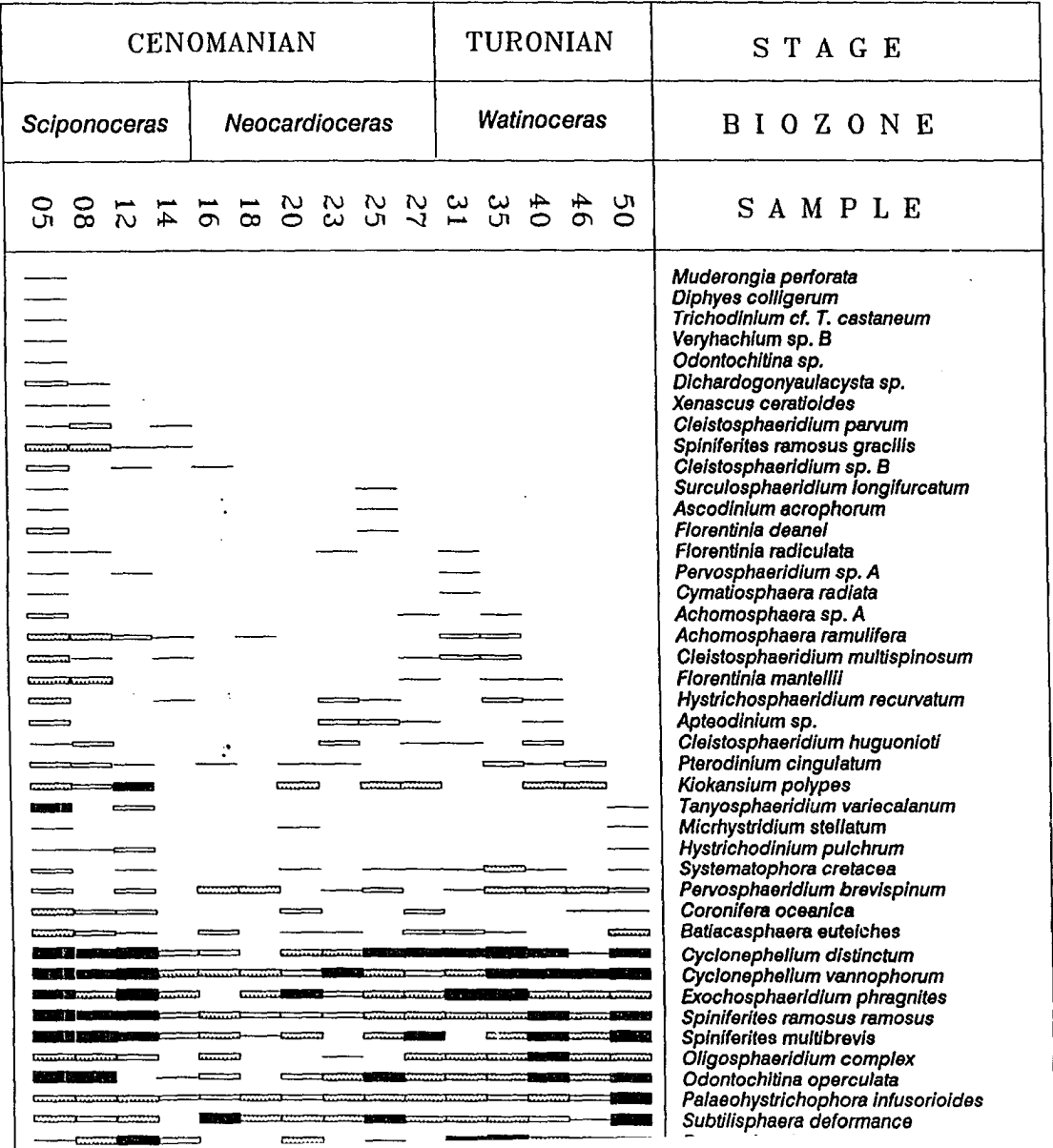
I. *Sciponoceras* Zone

The *Sciponoceras* zone yielded 54 species and subspecies (Fig. 15). Of them, eleven were restricted in this zone. The maximum species diversity and specimen abundance occur in the lowest part of the zone where the assemblage consists of 42 species and more than 400 individuals. Both species of *Cyclonephelium* and *Spiniferites* are the dominant types of the assemblage, and remain abundant through most of the zone. Other common types include *O.operculata*, *E.phragmites*, *K.polypes* and *O.complex*. Both species diversity and specimen abundance decrease gradually from the middle to the top within the zone. Only 14 species with less than 60 specimens are reported near the *Sciponoceras*-*Neocardioceras* zonal boundary. Stratigraphically-useful species *Cleistosphaeridium parvum* has its last occurrence prior to the *Sciponoceras*-*Neocardioceras* zonal boundary.

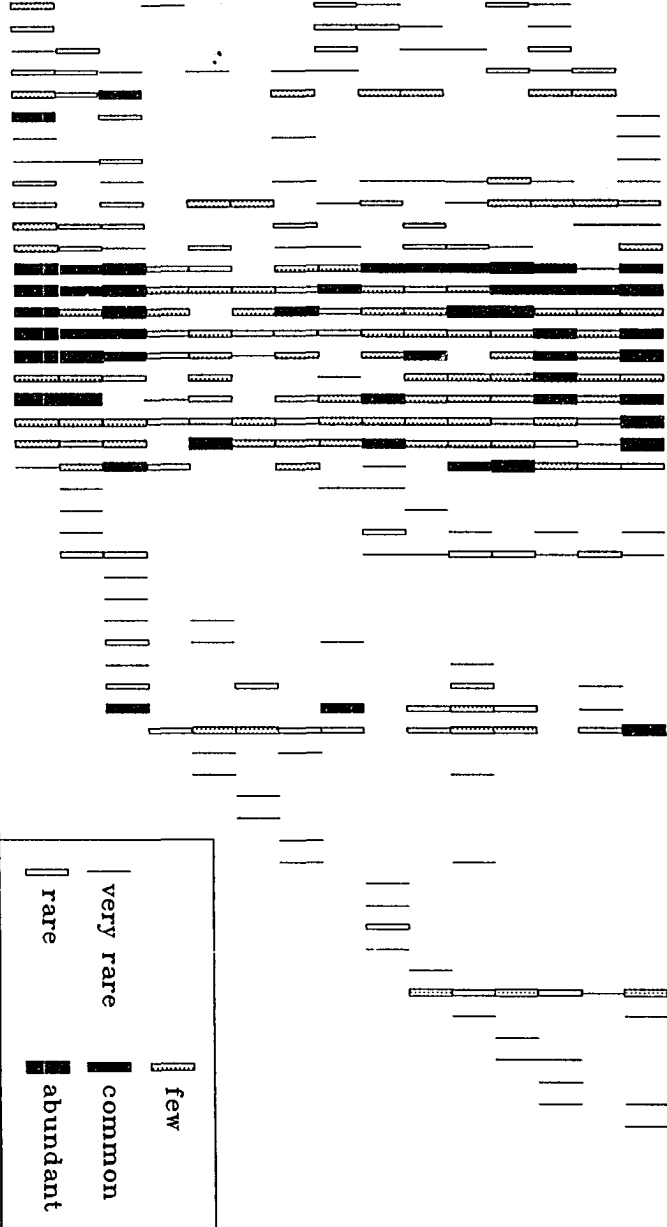
II. *Neocardioceras* Zone

Forty-nine species and subspecies were recorded in this

Fig. 15 Dinoflagellate-acritarch stratigraphy, Wahweap Wash, Utah (U1).



Hysitricosphaeridium recurvatum
Apteodinium sp.
Cleistosphaeridium huguonioti
Pterodinium cingulatum
Kiokansium polypes
Tanyosphaeridium variecalanum
Michystridium stellatum
Hystrichodinium pulchrum
Systematophora cretacea
Pervosphaeridium brevispinum
Coronifera oceanica
Batiacasphaera euteiches
Cyclonephellium distinctum
Cyclonephellium vannophorum
Exochosphaeridium phragmites
Spiniferites ramosus ramosus
Spiniferites multibrevis
Oligosphaeridium complex
Odontochitina operculata
Palaeohystrichophora infusorioides
Subtilisphaera deformance
Pervosphaeridium pseudohystrichodinium
Avellodinium sp.
Veryhachium sp. A
Odontochitina costata
Spiniferites sp. A
Luxadinium sp.
Xiphoniphorum alatum
Kallosphaeridium sp.
Heterosphaeridium sp.
Cyclonephellium sp.
Kickansium hesperum
Cyclonephellium membraniphorum
Subtilisphaera foliacea
Chytroelsphaeridia minor
Cleistosphaeridium sp. A
Gingiodinium evitti
Subtilisphaera pirmaensis
Pareodinia sp.
Michystridium sp. A
Ovoidinium verrucosum
Dinogymnium sp.
Apteodinium granulatum
Chytroelsphaeridia baetica
Florentinia cooksoniae
Spiniferites sp. C
Pterospermella cf. *P. pastelsi*
Spiniferites sp. B
Spiniferites ancoriferus
Spiniferites ramosus reticulatus
Veryhachium sp. C
Achomosphaera neptuni



Wash. Utah

zone (Fig. 15). Species diversity remains relatively low at the base (less than 20), and then gradually increase to middle 20 in the upper part of the zone. The main constituents of assemblages throughout the zone still are *C.distinctum*, *C.vannophorum*, *S.ramosus ramosus*, *S.multibrevis* and *E.phragmites*. Two peridinioid genera, *Palaeohystrichophora* and *Subtilisphaera*, which first appeared in the underlying intervals, become increasingly common upwards. Relatively short-ranging forms *Ascodinium acrophorum*, *S.longofurcatum* and *Chytroeisphaeridia minor* have their last appearance in the middle or near the top of the zone. Species richness does not show any difference across the C-T boundary.

III. *Watinoceras* Zone

The *Watinoceras* zone(lower and middle parts) yielded 44 species and subspecies(Fig. 15). Only five species are restricted to this zone, four of which belong to *Spiniferites* genus. Relatively diverse assemblages (average more than 20 species) were recorded from the entire studied interval of the zone. Amongst the abundant species are *P.infusorioides*, *S.deformance*, *S.foliacea*, *C.distinctum*, *C.vannophorum*, *S.multibrevis*, *S.ramosus ramosus* and *Pervosphaeridium pseudohystrichodinium*.

IV. Discussion

The dinoflagellate assemblages in the U1 section are

similar to those in the A1 section, although overall species richness is less. More than 70 taxa were reported. Most species, including all dominant types, are recorded in the Black Mesa area. Amongst the most abundant are *C.distinctum*, *C.vannophorum*, *E.phragmites*, *S.ramosus ramosus*, *S.multibrevis* and *O.operculata*. Other common species include *K.polypes* and *C.membraniphorum* in the *Sciponoceras* zone, *S.deformance* in the *Neocardioceras* zone, and *P.pseudohystrichodinium*, *S.foliacea*, *S.deformance* and *P.infusorioides* in the *Watinoceras* zone.

Spiniferites-type cysts flourish in the lower part of the *Sciponoceras* zone. For rest of the section, the *Cyclonephelium* group generally outnumbers the *Spiniferites* group, even though the margin is relatively small in some intervals. The abundance of peridinoid cysts increases in the upper part of the *Watinoceras* zone, but the concomitant decrease in the abundance of both *Spiniferites* and *Cyclonephelium* cysts is not as significant as in the A1 section.

Most of the biostratigraphic datums of important species, such as the last occurrence of *L.siphoniphorum*, *T.suspectum* and *E.spinosa* and the first occurrence of *H.difficile*, *Chatangiella* spp. and *S.protrusa*, are not recorded in this section. However, both the last occurrence of *C.parvum* in the *Sciponoceras* zone, and the last occurrence of *S.longifurcatum*, *A.acrophorum* and *C.minor* in the *Neocardioceras* zone suggest an age not later than the latest Cenomanian.

The most diverse and abundant assemblage occurs at the

base of the section, but then decrease gradually upwards. The relatively pronounced drop in species diversity occurs adjacent the *Sciponoceras*-*Neocardioceras* zonal boundary and at the bottom of *Neocardioceras* zone (Fig. 16). However, more than half of the species occurring in the preceding *Sciponoceras* zone, including all abundant and common taxa, do not disappear near the boundary, but extend successively to the overlying *Neocardioceras* zone. Unlike the Arizona section, the UI section does not show any signal of decrease in species diversity adjacent the C-T boundary. On the contrary, species diversity increases across the C-T boundary, although only on a small scale. There is no major difference in composition between the *Neocardioceras* and *Watinoceras* assemblages. The S/C ratio shows a more fluctuating pattern than the species diversity. There are three major peaks in S/C ratio throughout the sequence. The first occurs at the basal *Sciponoceras* zone which is coincident with the largest number of species. The second occurs in the upper part of the *Neocardioceras* zone which might be correlated with the change in species diversity from relatively lower (less than 20) to moderate (middle 20). The last one in the middle and upper parts of the *Watinoceras* zone studied is associated with a series of constantly moderately-diverse assemblages.

Depositional Organic Matter

Amorphous debris is the dominant type of organic residue

throughout the sequence (Fig. 16). Sporomorphs are few in most samples. Inertinite particles only play a relatively important role in certain intervals, such as the *Sciponoceras-Neocardioceras* zonal boundary and the C-T boundary. Dinoflagellate assemblages are moderately to relatively high diverse with few fluctuations throughout the section. A decline in species diversity in the top of the *Sciponoceras* zone coincides with the increased concentration of inertinite. There is no change in dinoflagellate species diversity across the C-T boundary even though inertinite particles show an increase.

Environmental Interpretation

The study of dinoflagellate species richness, S/C ratio, and associated depositional organic facies in the U1 section shows a similar trend to the A1 section. This is interpreted to be influenced by marine transgression-regression cycles (Fig. 16).

The largest species richness and increased S/C ratio associated with abundant amorphous debris at the base of the section suggest a rapid marine transgression. This transgressive pulse is also indicated by a basal silty and sandy bed (transgressive lag sand), numerous bentonite beds (increased volcanic activity), and a shift of faunal appearance from mixed brackish and normal marine to entirely normal marine (Elder 1987b). It is followed by a significant

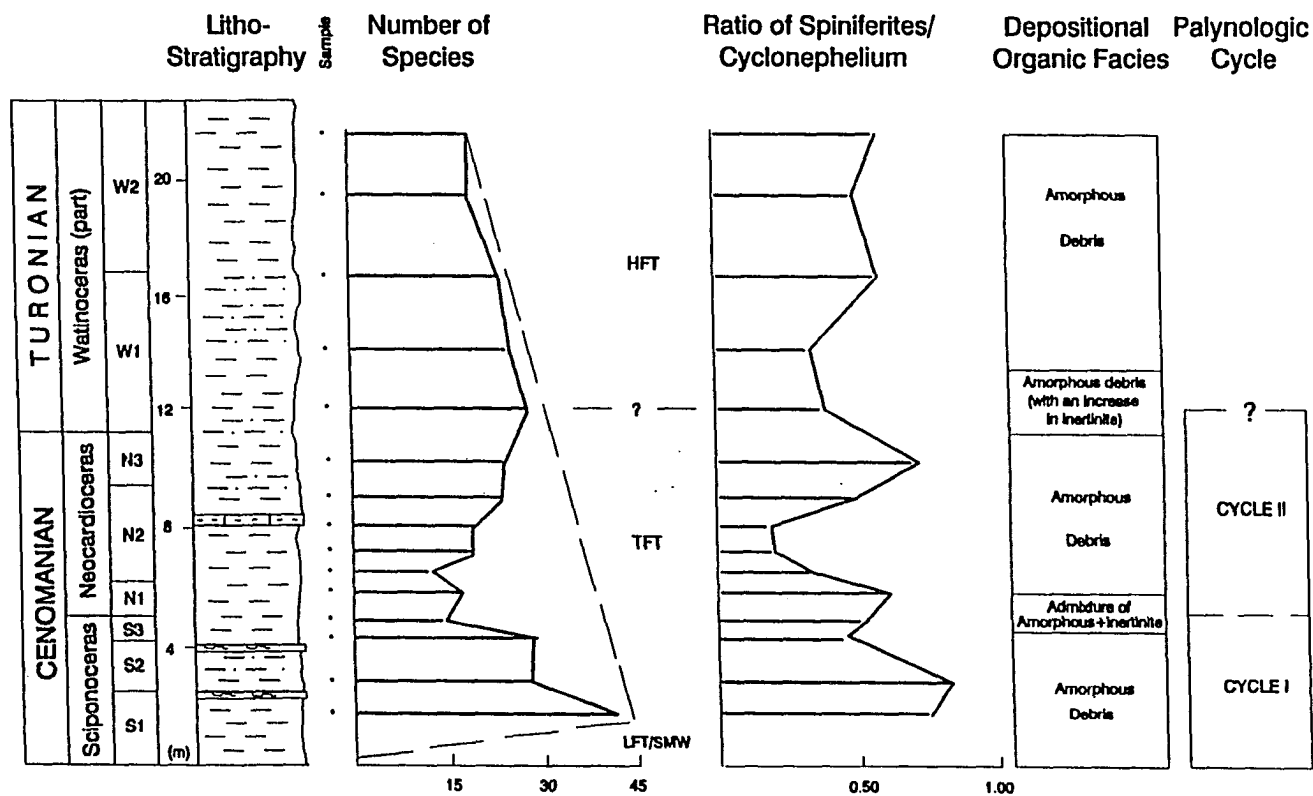


Fig. 16 Distribution of dinocyst species diversity, S/C ratio, and depositional organic facies in Wahweap Wash, Utah (U1).

decrease in species diversity associated with a decrease in specimen abundance, lower S/C ratios, and increased concentration of degraded land-plant materials and inertinites adjacent the *Sciponoceras-Neocardioceras* zonal boundary. All these palynologic evidences suggest a coastal depositional environment which was mostly formed during a regressive pulse.

The palynologic evidence of the second transgressive pulse in the lower part of the *Neocardioceras* zone includes an increase in both species diversity and S/C ratio and a decrease in the amount of terrigenous-origin organic matter. The marine flooding event is also indicated by an influx of diverse and abundant ammonite and inoceramid faunas above the basal silty bed of the *Neocardioceras* zone (Elder 1987b). However, a relatively smaller increase in species diversity and S/C ratio suggests that this transgressive pulse is less prominent than in the basal part of the preceding *Sciponoceras* zone. A possible smaller-scale regression adjacent the C-T boundary is indicated only by lower S/C values and a slight increase in concentration of inertinites.

The relatively diverse assemblages with higher S/C ratios defined by abundant amorphous fecal particles through the entire studied interval of the *Watinoceras* zone indicate a relatively stable open marine depositional environment. This interval is associated with an increase in carbonate content in the lower and middle parts of the zone (Elder 1987b).

In conclusion, species diversity, S/C ratio, and the

sequence of depositional organic facies in the U1 section show a similar marine transgression-regression cycle with that of the A1 section. The basal transgressive pulse is more easily recognized. The flourish of dinoflagellate cysts, both in terms of species diversity and specimen abundance, associated with abundant amorphous debris in the silty-beds of the basal *Sciponoceras* zone suggests a rapid transgression. It is followed by a regression adjacent the *Sciponoceras-Neocardioceras* zonal boundary. The regressive pulse is indicated by a marked decrease in species richness, lower S/C ratios, increasing amount of inertinite and other land-plant materials, and an increase in silt content (Elder 1987b). Compared with the A1 section, both regression and subsequent transgression across the *Sciponoceras-Neocardioceras* zonal boundary are less significant, since the changes in species diversity and in concentration of terrigenous organic detritus are relatively smaller. The C-T boundary is marked by no apparent decrease in species diversity, but only by a slight increase in the concentration of inertinites and the relatively lower S/C values. The real palynologic response to marine transgression-regression cycle may be hidden by the higher sedimentation rates. A stable open marine condition resulted from a continuous marine transgression is indicated by moderately to highly diverse dinoflagellate assemblages, relatively higher S/C ratios, and abundant amorphous debris through the entire studied interval of the *Watinoceras* zone.

This explanation is also supported by an increase in carbonate content and an influx of diverse epifaunal assemblage (Elder 1987b).

As shown in the A1 section, the largest number of species occurs at the beginning of marine transgression (lower part of the *Sciponoceras* zone), then decreases gradually from the upper part of the transgressive interval (*Neocardioceras* zone) into the followed highstand interval (*Watinoceras* zone). However, there is no apparent inflection point (Fig. 16) in the overall decreasing trend of species diversity which has been shown in the A1 section and the Bragg K-T boundary section (Habib et al., 1992). This seems to be related to the absence or the poor-development, if it exists, of a condensed interval during the *Neocardioceras*-*Watinoceras* zone time due to relatively high sedimentation rates (Elder 1987b). The lack of such a turning point makes it difficult to delimit the boundary between the transgressive and highstand intervals. The two sequence units (parasequence sets?) of the transgressive interval (*Sciponoceras* and *Neocardioceras* zone) are indicated by two palynologic cycles. The first cycle (*Sciponoceras* zone) is represented by the change in both species richness and the S/C ratio. The second (*Neocardioceras* zone) is only represented by the S/C ratio.

(iii) Dallas, Texas (T1)

The T1 section is situated on the northeastern edge of the Fort Worth Basin, and lies within the upper Britton Formation and lower Acadia Park Formation (Fig. 9). The environmental condition during the deposition of studied strata was a basically shallow marine shelf, even though it might be 200 km or more away from the eastern shoreline (Elder 1987b). The studied interval was characterized by the various deposition rates which were very high during the *Sciponoceras* zone time probably due to delta formation, and much lower in the *Watinoceras* zone (Elder 1987b).

Lithology

The *Sciponoceras* zone (only upper part) is characterized by dark green to green black, moderately-well laminated, noncalcareous to slightly-calcareous shales and claystones containing several limestone concretions. The lower and middle parts of the zone are not included in this study due to the great thickness of the *Sciponoceras* zone (greater than 40 m), and the lack of extensive and definitive marker-beds. The strata of the 1.5-meter-thick *Neocardioceras* zone are same as those of the underlying *Sciponoceras* zone. A disconformity around the C-T boundary is represented by a dramatic lithologic change and a lack of the earliest Turonian macrofaunas (Elder 1987b). Both organic carbon content and carbonate content increase abruptly across the C-T boundary

(Elder 1987b). The *Watinoceras* zone (only 2.5 meter in thickness) is characterized by olive green, moderately well indurated, calcareous shales and marly shales. The truncation at the top by the Camp Ranch Flagstone beds(*Collignonoceras woollgari* zone) indicates an interval of erosional removal or nondeposition (Elder 1987b).

Most samples, 11 out of 16, were collected from the upper part of the *Sciponoceras* zone. Five other include two from the *Neocardioceras* zone, and three from the *Watinoceras* zone.

Dinoflagellate Stratigraphy

I. *Sciponoceras* Zone

This zone yielded 50 species and subspecies of dinocysts and acritarchs(Fig. 17). Among them, more than half (30) are restricted in the *Sciponoceras* zone. *C.vannophorum*, *C.distinctum*, *Batiacasphaera euteiches*, *Chytroeisphaeridia minor*, *S.ramosus ramosus* and *S.multibrevis* are common. Stratigraphically useful species, such as *S.longifurcatum* and *Gingiodinium spinulosum*, have their last occurrences in the middle upper part of the zone. Relatively diverse assemblages are reported through most of the *Sciponoceras* zone except one pronounced drop in species richness at the base of Unit 19 (sample 21), approximately four meters below the *Sciponoceras-Neocardioceras* zonal boundary. Specimen abundance in all assemblages is low.

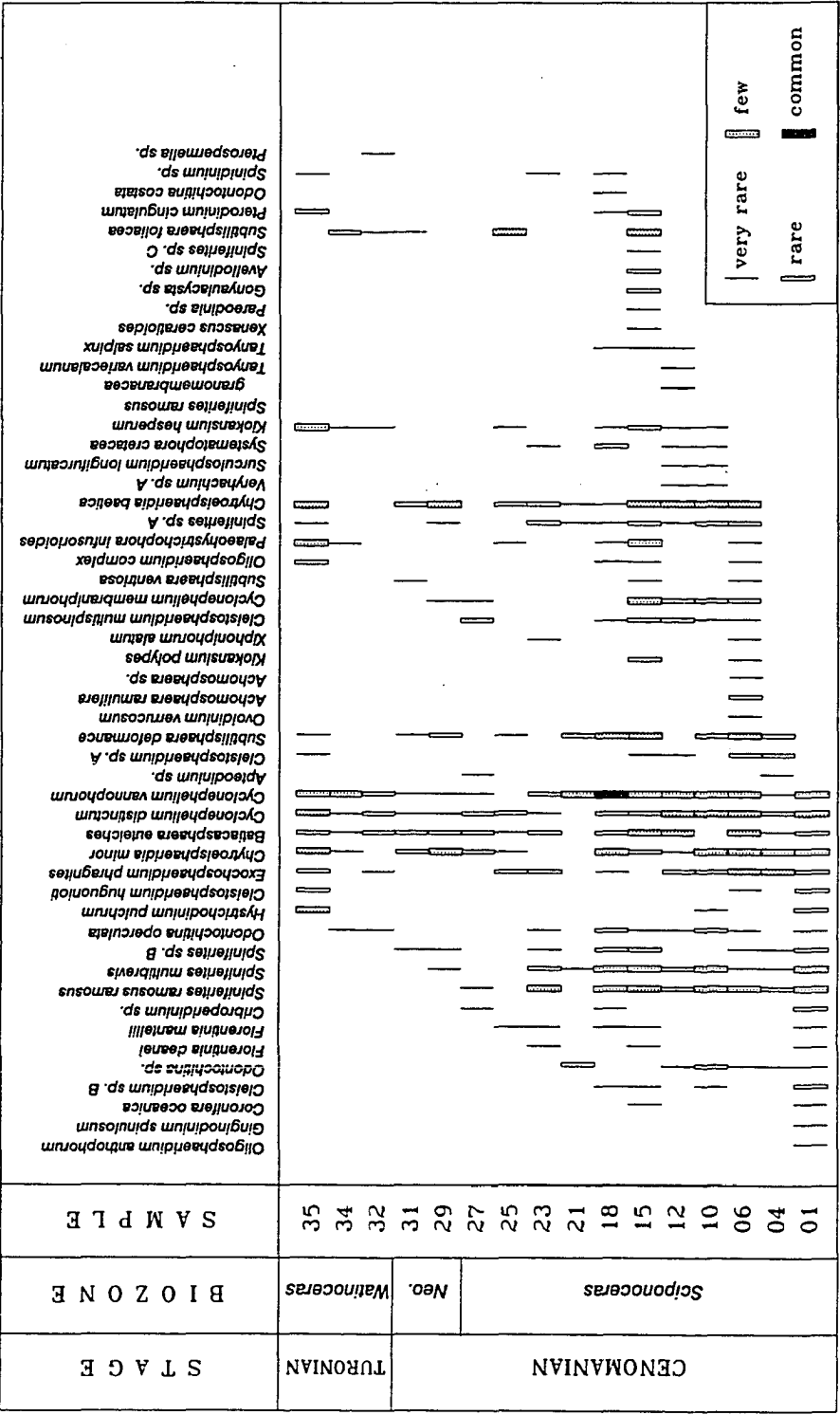


Fig. 17 Dinoflagellate-acritarch stratigraphy, Dallas, Texas (T1).

II. *Neocardioceras* Zone

This zone yielded only 12 species and subspecies that are all successively recorded in the underlying zone (Fig. 17). Amongst the common types are *Chytroeisphaeridia baetica*, *C. minor*, *B. euteiches*, *S. deformance* and *Cyclonephelium* spp.

III. *Watinoceras* Zone

The *Watinoceras* zone yielded 19 species and subspecies, 18 of which are recorded in the underlying *Neocardioceras* and *Sciponoceras* zones (Fig. 17). Species diversity and specimen abundance remain low in the basal and middle parts, and then increase slightly near the top of the section. *P. infusorioides*, *K. hesperum*, *Cyclonephelium* spp. and *Chytroeisphaeridia* spp. are relatively abundant and common types.

IV. Discussion

This section invariably contains many widely-distributed species, such as *C. distinctum*, *C. vannophorum*, *S. ramosus*, *O. operculata*, *E. phragmites* and *P. infusorioides*. However, several types which have not been reported as abundant or common in the western part of the seaway, including *C. baetica*, *C. minor* and *B. euteiches*, are also common through most the section.

The dominance in an assemblage by one or a few species is much lower than in the A1 or U1 sections. No single species

comprises more than 15% in any assemblage. The *Spiniferites* group is more abundant at the base than the rest of the section. However, the *Cyclonephelium* group generally outnumbers the *Spiniferites* group through the entire sequence. The increasing abundance of the peridinoid cysts in the lower Turonian *Watinoceras* zone is accompanied by no apparently concomitant decrease in specimen abundance of the *Spiniferites*- and *Cyclonephelium*-type cysts. Most of the section is also characterized by an unusual abundance of *B.euteiches* and species of *Chytroeisphaeridia*. Both *Batiacasphaera* and *Chytroeisphaeridia* were considered to be proximate cysts with ornamentation features of low relief (Stover and Evitt 1978). The major difference between those species and the *Cyclonephelium*-type cysts is the relative length of the projections. In contrary to the chorate *Spiniferites*-type cysts with long and complex processes, these proximate cysts with variable length of projections are more closely related to each other morphologically. Thus, they should have a same or similar ecological significance, although both *Chytroeisphaeridia* and *Batiacasphaera* have not been used as an environmental indicator before.

The important multiple occurrence datums mentioned in the standard profile (Chen et al., 1988) are not reported in this section. However, the last occurrence of *Surculosphaeridium longifurcatum* and *Gingiodinium spinulosum* in the upper part of the *Sciponoceras* zone indicates an age not younger than

latest Cenomanian, since both species have only been recorded in the Albian and Cenomanian sediments (Cookson and Eisenack 1960; Davey 1966).

The bottom of the study interval is characterized by moderately diverse assemblages. Species richness decreases systematically through the rest of the section. Only one big drop in species diversity is observed in the upper *Sciponoceras* zone (basal Unit 19), where the number of species decreases from 24 to only six. This decrease is coincident with no marked change in lithology, sedimentation rate and oceanic circulation pattern in this interval (Elder 1987b). The reason for the decrease in species richness is unknown. Species diversity remains relatively lower through the *Neocardioceras* and the *Watinoceras* zones.

The curve of S/C ratio shows a very similar pattern with the species richness. The higher S/C ratios coincide with the higher number of species at the base of the study interval, and then both decrease at the top of the *Sciponoceras* zone. The *Neocardioceras* and the *Watinoceras* zone are characterized by relatively lower S/C ratios. The uncommon abundance of proximate cysts (*Batiacasphaera* and *Chytroeisphaeridia*) through most of the section, and their close relation to the *Cyclonephelium*-type cysts suggest that these two types of cysts should be taken into the consideration when the S/C ratio is used. The lower revised S/C ratios (after adding these two cysts) suggest a more nearshore condition.

Depositional Organic Matter

This section shows a pure amorphous debris facies through the entire sequence (Fig. 18). Compared with its western counterparts (A1 and U1), Dallas section yields well-preserved, but less diverse and abundant, dinoflagellate assemblages. Nonmarine components, including spores and pollen grains, vascular tissues and oxidized remnants of land plants, are very few throughout the section. A significant decrease in both species richness and specimen abundance near the top of the *Sciponoceras* zone is coincident with no apparent change in the concomitant depositional organic matter.

Environmental Interpretation

Palynologic investigation from the T1 section (Fig. 18) shows that most of the section is characterized by low to moderately diverse assemblages with the dominant mixture of amorphous fecal pellet debris. The curves of species diversity and S/C ratio show little or no cycles which reflect marine transgression and regression. Unusually higher sedimentation rates during early-middle *Sciponoceras* zone time may have affected the distribution of dinocysts and other organic particles in sediments. It remains unclear why a deltaic setting with high sedimentation rates contains a very low concentration of non-marine organic components. Study of some samples from the base of the *Sciponoceras* zone would be

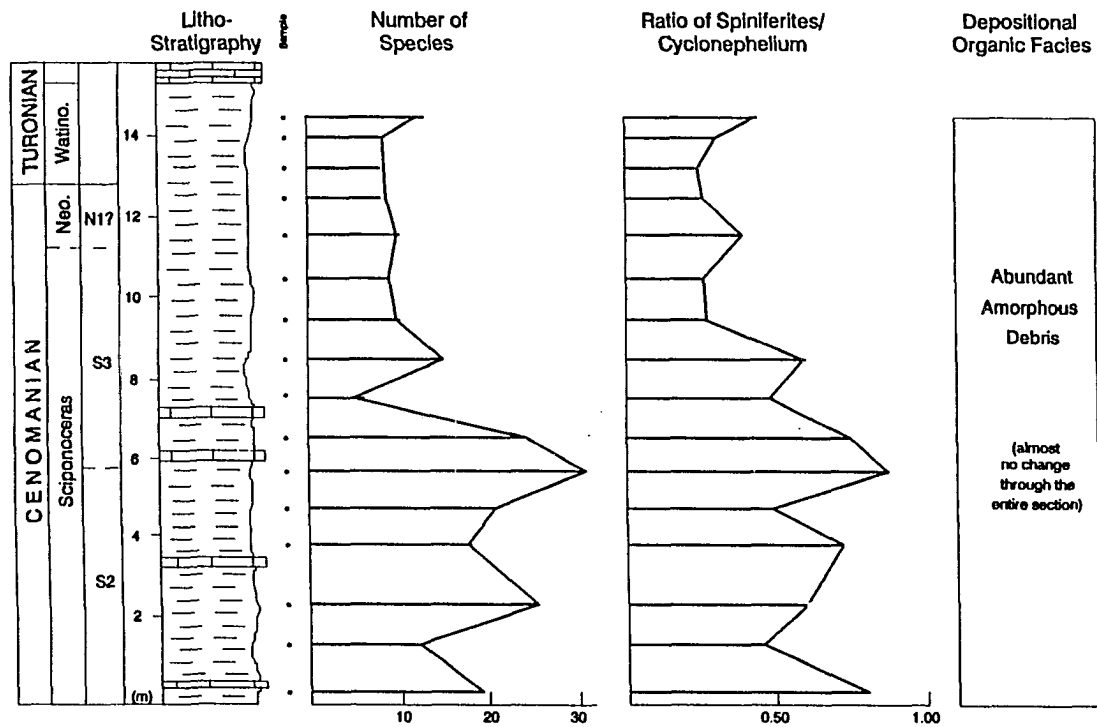


Fig. 18 Distribution of dinocyst species diversity, S/C ratio, and depositional organic facies in Dallas, Texas (T1).

helpful to see how species diversity and S/C ratio changed during the early period of marine transgression. Finally, a complete palynologic study is hampered by disconformities below and above the *Watinoceras* zone.

The base of the study interval (middle to upper parts of the *Sciponoceras* zone) is marked by a moderately diverse assemblage within an amorphous debris facies. This is followed by several fluctuations in species richness. The moderately diverse assemblages and medium to high S/C ratios associated with abundant amorphous debris and lower concentration of landplant materials indicate a generally open, relatively stable marine condition. The relatively significant drop in the number of species in the uppermost part of the zone is not associated with any significant change in depositional organic matter, sedimentation rate, carbonate content, and circulation pattern. Species diversity is recovered immediately above this horizon, but it never reaches the same level as below. Both the number of species and S/C ratio show little change through the rest of the section, reflecting a relatively stable shelf condition.

(iv) Clay-dominated Lithofacies Region Summary

The calcareous shales and claystones in the clay-dominated lithofacies region contain the most diverse and abundant dinoflagellate assemblages in the Western Interior basin.

In the western part of the basin, the distribution of dinoflagellates shows two apparent cycles in the late Cenomanian transgressive interval (*Sciponoceras* and *Neocardioceras* zone), which appears to be in response to the cyclic influence of small-scale marine transgression and regression. In contrast, cyclicity in the eastern clay-dominated lithofacies region (T1) is not as clear as in the western equivalent horizon due mainly to unusually high deposition rates. The cyclicity in the *Watinoceras* zone is very poor in both parts, indicating a relatively stable open marine condition throughout the basin during the highstand interval.

The first transgression across the entire basin at the basal *Sciponoceras* zone is characterized by the increase in species richness and in S/C ratio, associated with increased abundance of amorphous debris. This is consistent with the influx of the diverse *Sciponoceras* faunas above the basal transgressive lag sands (Elder 1987b). The largest number of species and the highest S/C ratio occur in the low to middle part of the early transgressive interval, reflecting a marked influence of the rapid expansion of the continental shelf at the beginning of the transgression. The significant decrease in species diversity and in S/C ratio, and the occurrence of abundant inertinite adjacent the *Sciponoceras*-*Neocardioceras* zonal boundary indicate a coastal condition possibly resulted from a marine regression. The disconformity and an increase in

silt content in this interval have been interpreted to be characteristics of the deposition during the period of a stillstand or a drop of sea level (Elder 1987b). The beginning of this drop in the upper *Sciponoceras* zone is approximately correlated with the macrofossil extinction step 1, and the final stage of this drop at the *Sciponoceras-Neocardioceras* zonal boundary with the macrofossil extinction step 2. These extinction steps were considered to be the result of regional environmental deterioration represented by higher sedimentation rates and decreased benthic oxygen concentrations (Elder 1987b). In contrast, dinoflagellates do not show any level of extinction at that time. Most dinoflagellate species temporarily disappear prior to the boundary, then reappear in the above strata.

The second pulse of sea level rise and marine transgression at the base of the *Neocardioceras* zone is suggested by multiple occurrence datums of dinocysts immediately above poorly fossiliferous samples. This is also supported by the immigration of the *Neocardioceras* zone faunas, condensation of strata and shift of the depocenter (Elder 1987b). The relatively diverse assemblages with medium to high S/C ratios remain unchanged in the rest of the zone until a decrease near the C-T boundary. The lower magnitude for the decrease in both species diversity (still no extinction) and S/C ratio, and lower abundance of inertinite, indicate that the C-T boundary event resulted from a possible

regression is not as great as that at the *Sciponoceras-Neocardioceras* zonal boundary. Elder's study (1987b) on macrofossils has indicated that regional environmental deterioration represented by decreased substrate stability may have contributed to the biotic extinctions (steps 3 to 5) in the *Neocardioceras* zone time. The unstable bottom condition is related to higher sedimentation rates in response to increased rainfall and fluvial input (Pratt 1984; Elder 1987b). The palynologic study shows there is indeed an increase in the amount of terrigenous organic detritus during this interval. However, landplant materials never constitute more than 10% of total organic matter.

The third transgression is not clearly indicated by species diversity and S/C ratio, but by the increase in carbonate content and in macrofaunal abundance (Elder 1987b). The moderately to highly diverse dinoflagellate assemblages, associated with the presence of abundant amorphous debris and the paucity of landplant materials, indicate a basically stable shelf environment.

In conclusion, the cyclicity of marine transgression and regression is better represented by palynologic evidence in the coastal and near-shore regions. The transgressive interval is indicated by the increased species diversity, relatively higher S/C ratios and abundant amorphous debris, whereas the regressive phase is suggested by the smaller numbers of species, lower S/C ratios, and the increased influence of

terrigenous organic detritus and inertinite. The lower part of the transgressive interval at the base of the *Sciponoceras* zone is marked by the most diverse assemblage with the highest S/C ratio. Then, both parameters decrease gradually through the upper part of the transgressive interval and the followed highstand interval. The pattern of change in species diversity during the late *Neocardioceras* and the early *Watinoceras* zone times seems to be related to sedimentation rate, and further to the development of a condensed interval. In the A1 section, the faster decrease in species diversity occurs at the time when a condensed interval is well developed (Elder 1987b). The absence of such a change in the U1 and T1 sections may be related to the poorly-developed condensed interval due to higher sedimentation rates. Detailed discussion will be in the following chapter.

2. Transitional Lithofacies Region

Three sections were studied in the transitional lithofacies region. They include one in the Henry Basin of southern Utah (U2); one in the Carthage-La Joya Basin near the eastern edge of the Rio Grande Depression (NM2); and another within south of the Souix Uplift on the eastern side of the Seaway (N1). The sections in this region display a transitional character from clay-dominated facies at the base (*Sciponoceras* zone) to carbonate-dominated facies at the top (*Neocardioceras* and *Watinoceras* zones). The delimitation of

lithofacies is therefore somewhat subjective. For instance, the base of the U2 section may be placed in the clay-dominated lithofacies region, whilst the middle and upper parts may be in the transitional lithofacies region (Elder 1987b).

(i) Egg Nog, Utah (U2)

The studied interval at the Egg Nog section (U2) includes the basal Tununk Shale Member of the Mancos Shale (Fig. 9) in the Henry Basin. The distance from the western margin of the seaway have changed from about 100 km at the base to more than 200 km at the top of the section. In general, the strata in this sequence contain a similar carbonate content to those in the comparable horizon of the U1 section (clay-dominated lithofacies region) (Elder 1987b). The U2 section was placed in the transitional lithofacies region based only on the development of a characteristic bivalve biofacies in the upper part of the *Neocardioceras* and *Watinoceras* zones (Elder 1987b).

Lithology

The basal part of the section is a layer of well-indurated, silty, non-calcareous sandstone which was thought to be a typical marine, transgressive lag deposit (Elder 1987b). Above this bed, the strata change continuously from sandy, calcareous siltstones (upper *Sciponoceras* zone-lower *Neocardioceras* zone), to clayey, calcareous siltstones (middle to upper *Neocardioceras* zone). The decrease in sand and silt

content associated with an increase in clay content upsection indicate continued transgression and increased water depth (Elder 1987b). Carbonate content increases sharply above the C-T boundary, indicating an increased planktonic productivity and/or decreased sedimentation rate (Elder 1987b). The *Watinoceras* zone is dominated by well-indurated, olive gray, slightly silty, highly calcareous claystone (Elder 1987b).

Seven samples were studied, including three from the *Neocardioceras* zone and four from the *Watinoceras* zone. No samples were collected in the *Sciponoceras* zone because of its high sand content which is unfavorable to the preservation of dinocysts.

Dinoflagellate Stratigraphy

No detailed discussion for each biozone will be conducted because the samples are only from the *Neocardioceras* and *Watinoceras* zones, and the number of samples from each biozone is low (Fig. 19).

The dinoflagellate assemblages are similar to those in the U1 section, except less diverse. Among 30 species and subspecies reported, the species of *Cyclonephelium*, such as *C.vannophorum*, *C.membraniphorum*, *C.distinctum*, and *S.multibrevis* are the most abundant types. Other common species include *E.phragmites*, *S.amosus amosus*, *K.polypes* and *P.infusorioides*.

No single or few taxa dominate the assemblage. However,

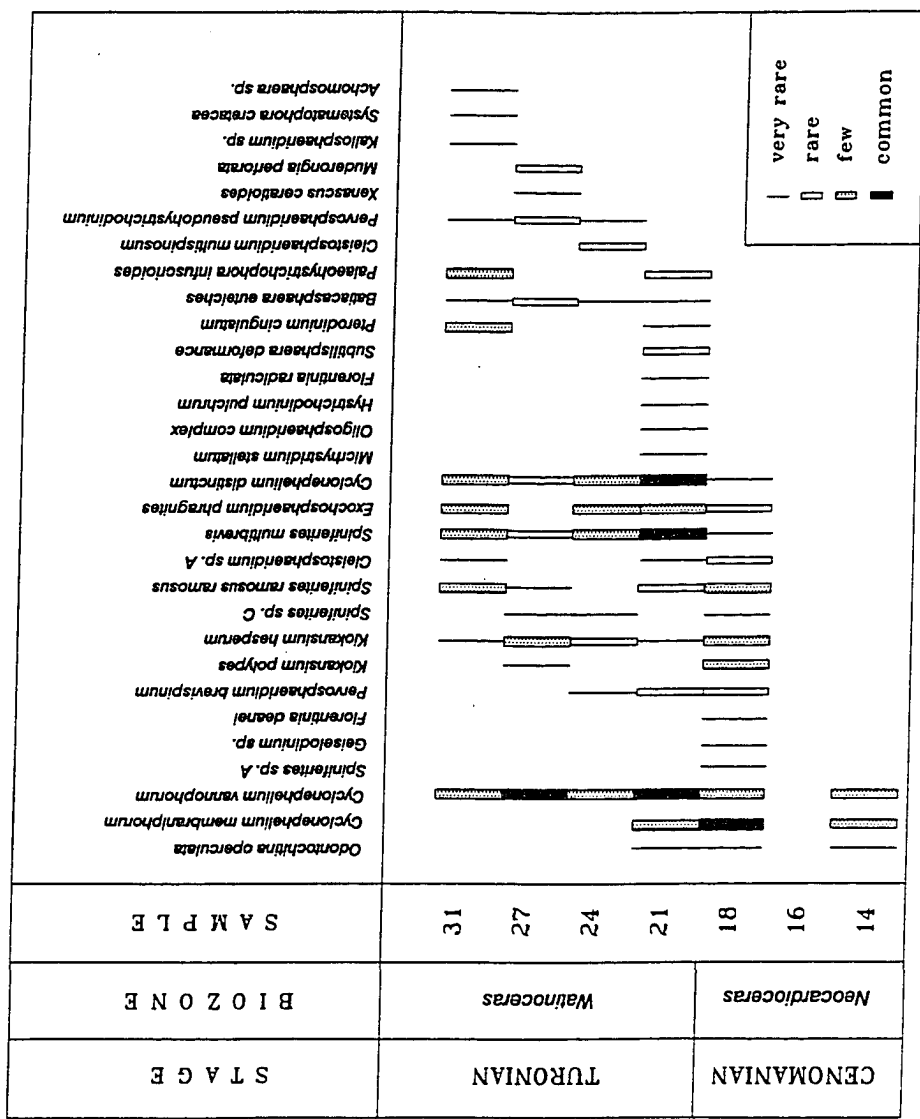


Fig. 19 Dinoflagellate-acritarch stratigraphy, Egg Nog, Utah (U2).

the *Cyclonephelium*-type cysts are apparently more abundant than the *Spiniferites*-type cysts at the basal section (*Neocardioceras* zone). The *Spiniferites* group increases its abundance in the lower part of the *Watinoceras* zone, but it never surpasses the *Cyclonephelium* group. The peridinioid *P.infusorioides* becomes more common in the upper part of the *Watinoceras* zone, along with a single species of *Subtilisphaera* (*S.deformance*).

Species richness increases upwards from one to three species at the base of the study interval (middle part of the *Neocardioceras* zone) to 15 in the upper part of the *Neocardioceras* zone. The maximum species diversity (18) occurs at the basal *Watinoceras* zone. The rest of the *Watinoceras* zone is characterized by moderately diverse assemblages. Species diversity in the U2 section exhibits a similar pattern to the U1 section where it increases slightly across the C-T boundary.

Depositional Organic Matter

A constant organic facies, comprised mostly of amorphous debris along with inertinite and terrigenous organic detritus, is found throughout the section (Fig. 20). The dinoflagellate assemblages are similar to those in the Wahweap Wash section (U1), except less diverse and less abundant.

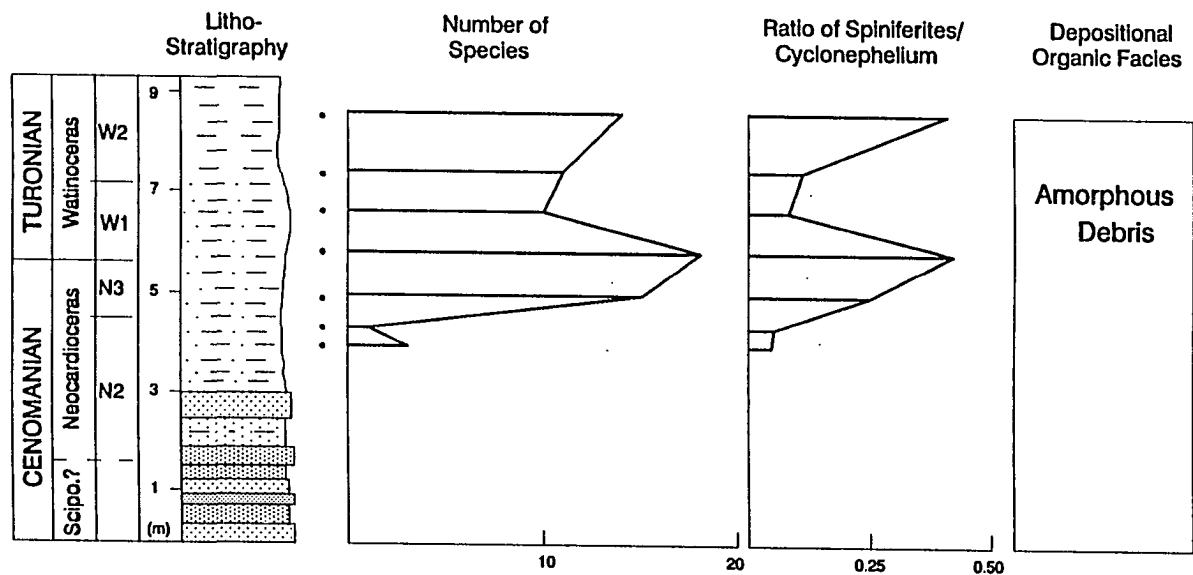


Fig. 20 Distribution of dinocyst species diversity, S/C ratio, and depositional organic facies in Egg Nog, Utah (U2).

Environmental Interpretation

The basal sandstone (possible marine transgressive lag deposit), and lowest three meters of section were not sampled due to the relatively high sand and silt content. The samples from the middle to upper part of the *Neocardioceras* zone (Unit 14 and 15) contain assemblages of very low diversity and low abundance (Fig. 20). This seems not to relate to higher sand content, since both units contain little or no silt. The significant amounts of bentonite in the sediments may be attributed to the low concentration of dinocysts. Both species diversity and numerical abundance increase in the upper part of the *Neocardioceras* zone, and remain the same level through the rest of the section (lower part of the *Watinoceras* zone). The association of moderately diverse dinoflagellate assemblage with abundant amorphous debris suggests a normal marine environment during the period of a sea level highstand, which has been supported by increased carbonate content (Elder 1987b). The curve of species diversity in the *Watinoceras* zone shows a same pattern as in the A1 and U1 sections that a general reduction in the number of species reflects shoreline progradation in the highstand interval. Lithologically and palynologically, this section is better placed in the clay-dominated lithofacies region.

(ii) Carthage, New Mexico (NM2)

The section studied at Carthage (NM2) lies in the Bridge Creek Limestone Beds of the Mancos Shale (Fig. 9). The distance from the western shoreline of the seaway at the time of deposition of strata was estimated to be approximately 250-300 km (Elder 1987b). The relatively complete work of stratigraphy and paleontology (mainly macrofossils) of this section has been done by Elder (1987b, 1989).

Lithology

The basal *Sciponoceras* zone is marked by a well-indurated, olive gray, marlstone bed (LS1). The rest of the *Sciponoceras* zone is characterized by moderately well-indurated, olive gray, slightly silty, calcareous shales interbedded with two limestone concretion beds (LS3 and LS4). The unusual thicker lower part of the zone, compared with the stratigraphic equivalent at other localities, was thought to be the result of downslope sediment reworking and bypass off the Tongue of the Dakota Sandstone to the north and west (Elder 1987b, 1991). The lower part of the *Neocardioceras* zone is characterized by moderately well-indurated, dark olive gray, slightly silty, calcareous shales and marly shales interbedded with thin well-indurated marlstones (LS5 and LS6). The upper part of the zone consists of a similar association of rocks, except little condensed. The lower part of the *Watinoceras* zone is also somewhat condensed. The rest of the zone is

characterized by moderately well- to well-indurated, dark olive gray, marly shales interbedded with three gray marlstone beds (LS8, LS9, and LS10) (Elder 1987b).

Organic carbon content is almost same (around 0.5%) throughout the section except the top bed (Unit 61). Carbonate content in the calcareous shales increases upsection (calcareous shales to marly shales), suggesting an increase in planktonic productivity (Elder 1987b). An apparent decrease in carbonate content is associated with a higher organic carbon content (>1%) in the top of the *Watinoceras* zone. This may be due to increased sedimentation rates, which lead to a dilution of carbonate content and a better preservation of organic matter through a rapid burial (Elder 1987b).

Eighteen samples were collected, including seven from the *Sciponoceras* zone (5.5-meter-thick), five from the *Neocardioceras* zone (2.5-meter-thick), and six from *Watinoceras* zone (5-meter-thick). Most samples in this study are dark olive gray, slightly silty, calcareous shales.

Dinoflagellate Stratigraphy

I. *Sciponoceras* Zone

Ninty-five species and subspecies of dinocysts and acritarchs were recorded (more than 80% of the total taxa), 33 of which were restricted to this interval (Fig. 21). This zone contains the same or similar dinocyst composition as in the

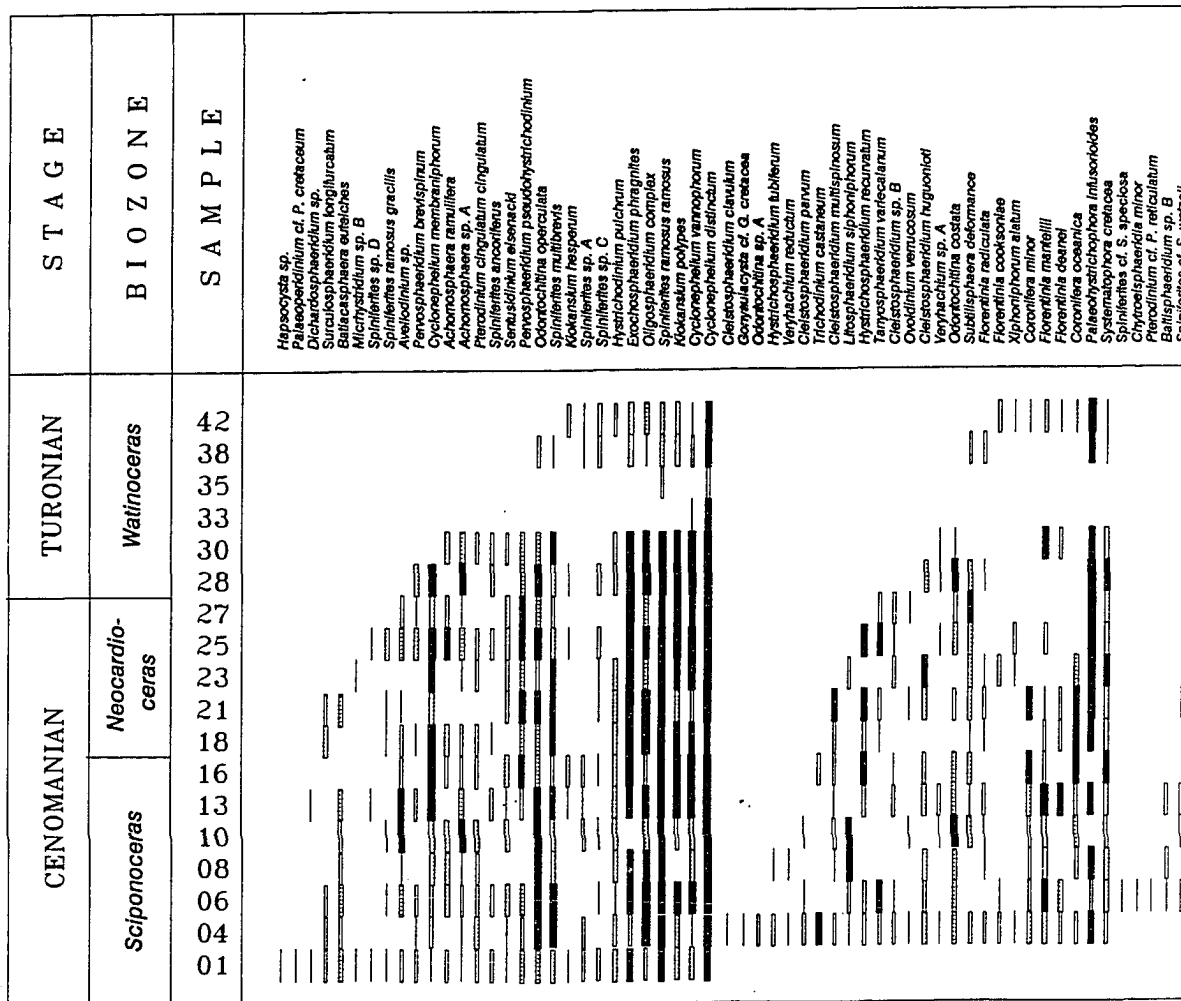


Fig. 21 Dinoflagellate-acritarch stratigraphy, Carthage, New Mexico (NM2).

Cleistosphaeridium parvum
Tripos sp.
Cleistosphaeridium multiflosum
Leiosphaeridium siphonophorum
Hydractinia sp.
Teniosphaeridium varicellatum
Cleistosphaeridium sp. B
Ovalidium verrucosum
Cleistosphaeridium huguonoidi
Verrucium sp. A
Odontochitina costata
Subtilisphaera deformans
Florentinia radiculata
Xiphonophora alatum
Coronilla minor
Florentinia maritima
Florentinia oceanii
Coronilla costata
Subtilisphaera crocata
Spiliferites cf. S. apocosa
Hydractinia minor
Pterodinium cf. P. reticulatum
Baliosphaeridium sp. B
Spiliferites cf. S. wetzelii
Paradonia sp.
Microtyridium sp. A
Achomospaera sagena
Cleistosphaeridium sp. A
Teniosphaeridium salpinx
Spiliferites sp. B
Isabelidium magnum
Gonyaulacysta sp. A
Verrucium sp. B
Bourdinium psidium
Spiliferites sp. E
Achomospaera sp. B
Achomospaera repum
Sentadidium sp.
Subtilisphaera foliacea
Dinocyst Form M
Pterodinium alloum
Chrysa cf. D. colligum
Flabellidium
Baliosphaeridium sp. A
Verrucium sp. C
Pterospermella harti
Cleistosphaeridium oculare
Gonyaulacysta sp. B
Gonyaulacysta cassidata
Coronilla sp.
Spiliferites ramosus granomembraceus
Chytosphaerella baetica
Microtyridium sp. B
Pterodinium lingulatum subsp.
Ascodinium parvum
Subtilisphaera portis-maritae
Pterosphaeridium sp. A
Operculodinium sp.
Spiliferites sp. F
Pterosphaeridium sp. B
Pterosphaeridium sp. C
Achomospaera sp. C
Spiliferites ramosus reticulatus
Subtilisphaera primaensis
Geleodinium sp.
Baliosphaera cf. B. reticulata
Spiliferites paradoxus
Xiphonophora sp.
Subtilisphaera senegalensis
Pterospermella harti
Odontochitina sp. B
Spiliferites cornutus laevimurus
Cordosphaeridium sp.
Oligosphaeridium cf. O. reticulatum
Dinocyst sp. F
Moderongia perforata
Catatomocystis ? sp.

| very rare | common
 | rare | abundant

Arizona section. Several long-ranging, widely-distributed species, such as *S.ramosus ramosus*, *S.multibrevis*, *C.distictum*, *O.operculata*, *O.costata*, *C.membraniphorum*, *K.polypes*, and *E.phragmites*, are the most abundant and extend through the *Sciponoceras-Neocardioceras* zonal boundary. *Cleistosphaeridium parvum* and *Bourkidinium psilatum* have their last appearance in the upper part of the zone. Other species which can be an age-determination indicator include *S.longifurcatum*, *L.siphoniphorum* and *Achomosphaera sagena*. There is no significant change in species diversity and numerical abundance within the zone and across the *Sciponoceras-Neocardioceras* zonal boundary.

II. *Neocardioceras* Zone

This zone yielded 77 species and subspecies of dinocysts and acritarchs (Fig. 21). Twelve taxa first appear at this interval. Of them, seven are restricted to this zone. The other 65 taxa are reported through the *Sciponoceras-Neocardioceras* zonal boundary. The assemblages in lower part are dominated by *O.complex*, and to a lesser extent by species such as *E.phragmites*, *C.distinctum*, *S.ramosus ramosus* and *P.pseudohystrichodinium*. The upper part is characterized by abundant *S.ramosus ramosus*, *C.distinctum*, *K.polypes*, *E.phragmites* and *P.infusorioides*. Most of the abundant types in this zone are also recorded successively in the overlying zone. *L.siphoniphorum*, *S.longifurcatum*, *Ascodinium parvum* and

A.sagena have their range tops within the middle or upper part of the *Neocardioceras* zone.

III. *Watinoceras* Zone

The main constituents in the lower part of the *Watinoceras* zone are similar to those in the underlying zone (Fig. 21). Amongst the most abundant are *E.phragmites*, *C.distictum* and *K.polypes*. *S.ramosus ramosus*, *O.complex*, *C.vannophorum*, *O.operculata*, *Systematophora cretacea* and *P.infusorioides* are very common. A pronounced decrease in species diversity and numerical abundance occurs in the middle part of the zone (Unit 51 and 53) where assemblages contain only two to three species with 10-15 specimens. This is in a big contrast with the underlying assemblages consisting of 27-30 species and more than 300 specimens. Moderately diverse assemblages are recorded in the upper part of the *Watinoceras* zone. The dominant types are *P.infusorioides* and *C.distinctum*, with *S.ramosus ramosus* and *Kiokansium* spp. common. Most species (16 out of 19) are from the underlying assemblages, only three species first appear in this interval.

IV. Discussion

Dinoflagellate assemblages in the Carthage section are similar to those in the Black Mesa section in both species diversity and specimen abundance. Most samples contained 200 or more individuals, and totally more than 100 species and

subspecies were recorded throughout the section. The *Spiniferites*-type cysts, such as *S.ramosus ramosus*, *S.multibrevis* and *O.complex* are more abundant than those in the Arizona section, e.g., *O.complex* is the most common type in the lower part of the *Neocardioceras* zone. *E.phragmites*, *C.distinctum* and *S.ramosus ramosus* are equally abundant in the upper part of the *Neocardioceras* zone. Amongst the abundant taxa in the lower part of the *Watinoceras* zone are *K.polypes*, *E.phragmites*, *S.ramosus ramosus* and *C.distinctum*. The peridinioid *P.infusorioides* is common throughout the *Neocardioceras* and *Watinoceras* zones.

As discussed above, this section contains more abundant *Spiniferites*-type cysts than the A1 section, although *Cyclonephelium*-type cysts are still main components in most intervals. The lower part of the section (most of the *Sciponoceras* and *Neocardioceras* zones) is characterized by a relatively well-balanced *Cyclonephelium* and *Spiniferites* association. The *Cyclonephelium* group remains abundant, but it never becomes as dominant in the assemblages as it is in the clay-dominated regions. The *Spiniferites* group is also abundant, and even outnumbers the *Cyclonephelium* group at the base of the *Sciponoceras* and *Neocardioceras* zones. The upper part of the *Watinoceras* zone is characterized by the higher percentages of the peridinioid cysts with a gradual decrease in the abundance of the *Cyclonephelium* and *Spiniferites* groups.

Many of the occurrence datams of the important species as listed by Chen et al. (1988) are not found in the NM2 section. However, the last occurrence of *C.parvum* and *B.psilatium* in the *Sciponoceras* zone, and *S.longifurcatum*, *A.parvum* and *A.sagena* in the *Neocardioceras* zone suggests a late Cenomanian age since these species have only been reported to be present in the European Cenomanian strata (Davey et al., 1966; Davey 1969).

In comparison with the A1 section, the fluctuation of species diversity throughout the NM2 section is relatively small, especially adjacent the *Sciponoceras-Neocardioceras* zonal boundary and the C-T boundary. Most common taxa appear through the entire section except in the middle part of the *Watinoceras* zone, where more than 90% of the species disappear temporarily. The species richness recovers following the drop, but it still lower than those in the preceding strata.

Depositional Organic Matter

This section is characterized by a typical amorphous debris facies which includes a well-preserved, diverse and abundant dinoflagellate assemblage, low concentration of inertinite and landplant materials (Fig. 22). Inertinite particles show a slight increase near the *Sciponoceras-Neocardioceras* zonal boundary and the C-T boundary, which is not coincident with the decrease of species diversity as shown in the Arizona sequence. The significant decrease in species

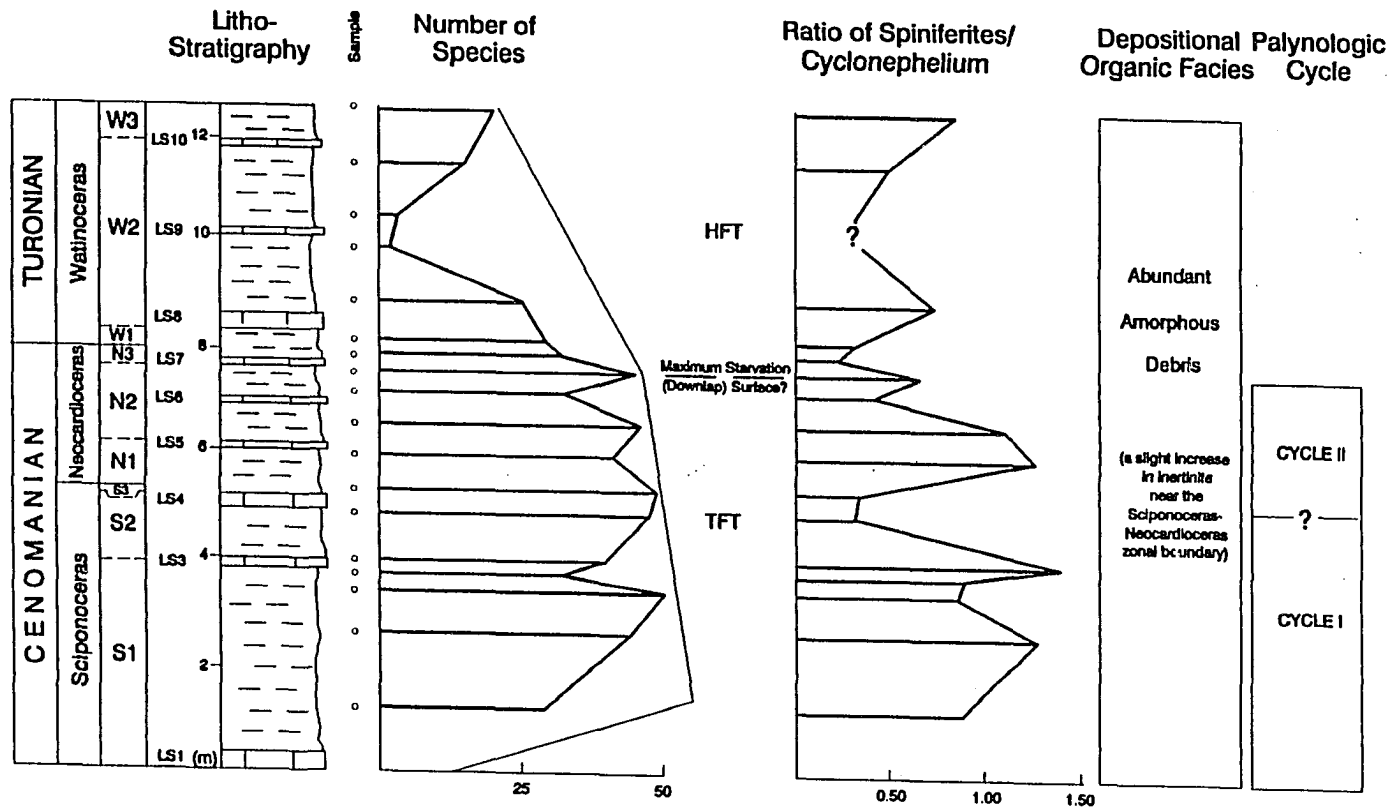


Fig. 22 Distribution of dinocyst species diversity, S/C ratio, and depositional organic facies in Carthage, New Mexico (NM2).

richness (27 to three species) only occurs in the middle part of the *Watinoceras* zone associated with no apparent change in depositional organic facies.

Environmental Interpretation

Palynologic data from the NM2 section (Fig. 22) show that most of the study interval is characterized by diverse dinoflagellate assemblages, associated with abundant amorphous debris. Less and smaller fluctuations in species richness and in the associated organic facies than in the clay-dominated A1 section suggest a relatively stable depositional environment which has been less influenced by the cyclicity of marine transgression and regression.

The base of the *Sciponoceras* zone contains a relatively diverse (29 species) dinoflagellate assemblage and a high concentration of fecal amorphous debris. This is followed by a gradual increase in species diversity in the lower and middle parts of the zone. The absence of an abrupt increase in species richness and the presence of a higher S/C ratio at the base of the section suggest that this area had been inundated for a period of time prior to deposition of sediments at the base of the study interval. This is supported by sedimentologic evidence that the basal *Sciponoceras* zone was not marked by a transgressive lag sand, but by a well-indurated marlstone bed (Elder 1987b). Species richness does not vary significantly across the *Sciponoceras*-*Neocardioceras*

zonal boundary. Only a pronounced drop in S/C ratio adjacent the *Sciponoceras*-*Neocardioceras* zonal boundary implies a change in depositional environments towards nearer-shore which is possibly in response to a small-scale regression. Relatively high-diversity assemblages through the entire *Sciponoceras* zone and most of the *Neocardioceras* zone indicate that depositional environments at that time are basically normal marine conditions during the period of increased transgression. A gradual change in species richness and in S/C ratio occurs in the condensed strata adjacent the C-T boundary. This shows a similar result with the finding of Habib and others(1992) in the Cretaceous-Tertiary Atlantic Coastal Plain that number of dinoflagellate species decreased from the transgressive systems tract through the condensed section during the maximum transgression (Habib et al., 1992). A significant decrease in species diversity occurs in the middle part of the *Watinoceras* zone(above the condensed interval). Compared with 27 species and more than 300 specimens in the preceding assemblage, only two to three species with less than ten specimens are recorded in this interval. The sharp decrease in the number of dinocyst species may be related not only to the increased shoreline progradation in the highstand interval, but also to the increased degree of retransportation and redeposition of dinocysts due to an increase in current activity indicated by an apparent ammonite influx and the development of the starved

calcarenite ripples (Elder 1987b).

In conclusion, dinoflagellates species richness and depositional organic facies in NM2 section reflect only part of a complete cycle of marine transgression-regression. The transgressive pulses at the base of the *Sciponoceras* zone and *Neocardioceras* zone, particularly the first one, are indicated not only by an increased species richness and higher S/C ratios, but also by a decreased silt content and an influx of diverse faunas. The lack of well-represented regressive intervals is due mostly to its locality relatively far away from the paleo-shoreline which has hardly been influenced by smaller-scale regression. Relatively diverse dinoflagellate assemblages associated with abundant amorphous debris in a well-indurated marlstone bed, rather than in a siltstone or silty claystone bed at the base of the section indicates that those strata were not deposited at the basal part of transgressive interval. The transgressive lag sand was reported 10 meter or more below the base of the *Sciponoceras* zone in the offshore sections, showing the marked diachroneity of the transgressive surface (Elder 1987b). The relatively large numbers of dinocyst species and unchanged amorphous debris matrix through most of the *Sciponoceras* and *Neocardioceras* zones suggest an open marine condition. The environmental change towards nearshore is only inferred by a marked reduction in S/C ratio adjacent the *Sciponoceras*-*Neocardioceras* zonal boundary.

The curves of species diversity in the NM2 section look like those found in Arizona and Alabama, even though the amplitude of variations is relatively lower. The number of dinoflagellate species increases in the lower part of the transgressive phase, then decreases more or less in the upper part of the transgressive pulse, and the following condensed section and highstand interval. The parasequence sets are indicated more by the cyclicity of S/C ratios, rather than the changes in species diversity and inertinite abundance in the *Sciponoceras* and *Neocardioceras* zones. An apparent inflection in the overall decreasing trend of species diversity is also found in the upper part of the *Neocardioceras* zone (a position slightly higher than that in the A1 section), which was thought to be within the upper part of the condensed interval (Elder 1987b). This is a little different from the finding of Habib and others in the Braggs area of Alabama, who showed that such an inflection point occurred just below the top of the condensed section (Habib et al., 1992).

(iii) Ponca, Nebraska (N1)

The Ponca section (N1) lies in the Greenhorn Formation, and lay probably 300 to 400 km west of the eastern margin of the seaway during deposition of strata in the study interval. In general, the carbonate contents of strata in this section show no difference with those in the sections of the basin center, such as the C2 section (Elder 1987b). This section was

placed in the transitional lithofacies region due mainly to the development of highly calcareous to marly shales in the upper Cenomanian and lower Turonian strata (Elder 1987b).

Lithology

The base of the section is characterized by moderately well-indurated, olive gray, silty, calcareous shales. The other parts of the section are well-indurated, olive gray marlstone. The rocks of the studied interval are all chalky (greater than 50% carbonate content). Organic carbon content is also high(1.8-5.4%), particularly in the late Cenomanian *Sciponoceras* and *Neocardioceras* zones. The unusually high level of organic carbon concentration may be the result of better preservation due to low benthic oxygen level (Elder 1987b).

Ten samples were studied, including four from the *Sciponoceras* zone(2.5-meter-thick), three from the *Neocardioceras* zone(1.0-meter-thick), and three from the *Watinoceras* zone(2.2-meter-thick).

Dinoflagellate Stratigraphy

Twenty-three species and subspecies of dinoflagellate cysts were recorded (Fig. 23). The most frequently occurring species is *C.distinctum* which appears in 7 out of 10 samples. This is followed by *C.vannophorum* (5 samples), *S.ramosus ramosus* (5 samples), *S.multibrevis* (5 samples), and

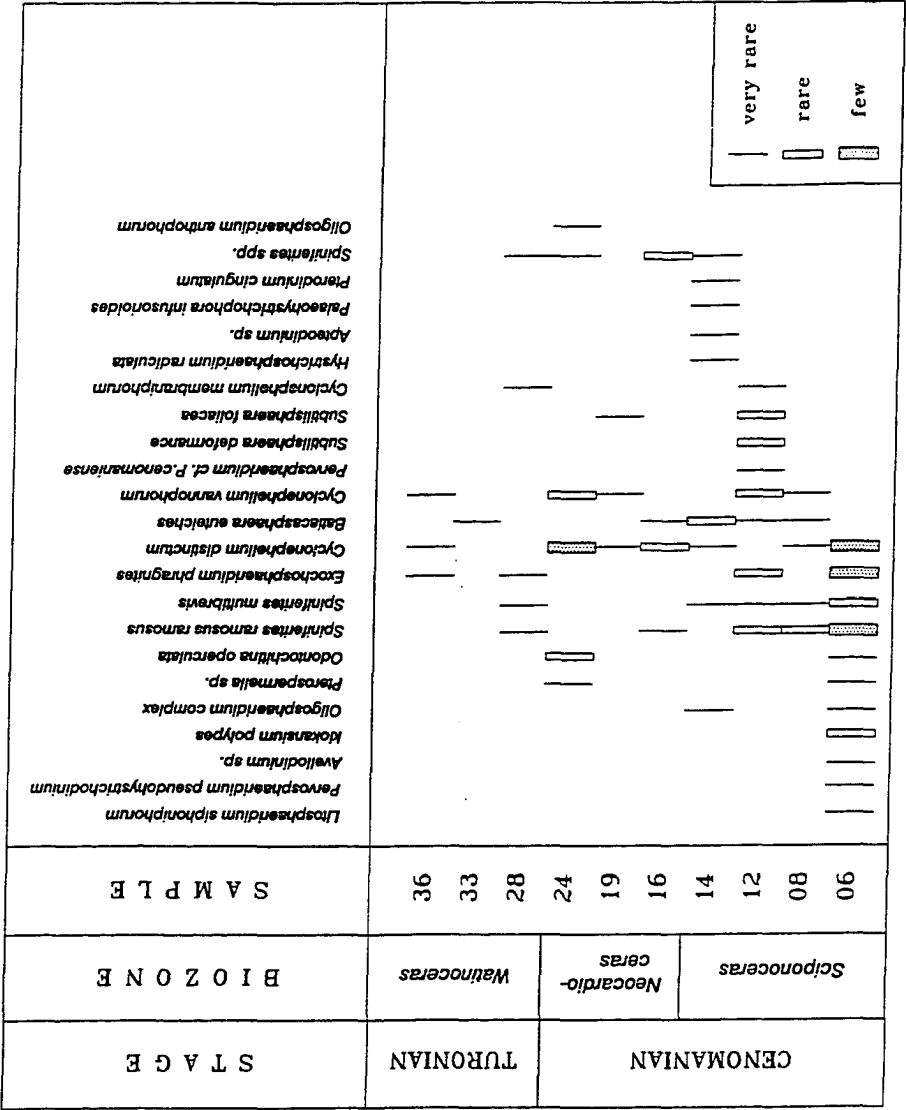


Fig. 23 Dinoflagellate-acritarch stratigraphy, Ponca, Nebraska (N1).

B.euteiches (5 samples). Each of the remaining 18 species occurs in fewer than five samples.

The last occurrence of *L.siphoniphorum* and *Pervosphaeridium cenomaniense* in the *Sciponoceras* zone suggests a late Cenomanian age.

Both the *Spiniferites* and *Cyclonephelium* groups are relatively abundant in most samples. No detailed S/C ratio analysis has been done due to the smaller specimen numbers.

The entire section is characterized by low-diversity assemblages. Species richness ranges from 11 (maximum) at the base of the section, gradually decreasing through the upper part of the *Sciponoceras* zone (5-9), and the *Neocardioceras* zone(3-6), finally into the *Watinoceras* zone(1-3).

Depositional Organic Matter

The N1 section consists almost entirely of amorphous debris (Fig. 24). Approximate 10% of such particles are relatively dark in color which suggest a relatively higher degree of oxidization and degradation which might have occurred within the surface water column or through descending route to the bottom of the basin. The concentration of inertinites and land plant materials is very low. Dinoflagellate assemblages are low diversity in most samples, and only contain a single or few specimens in the upper part of the *Watinoceras* zone.

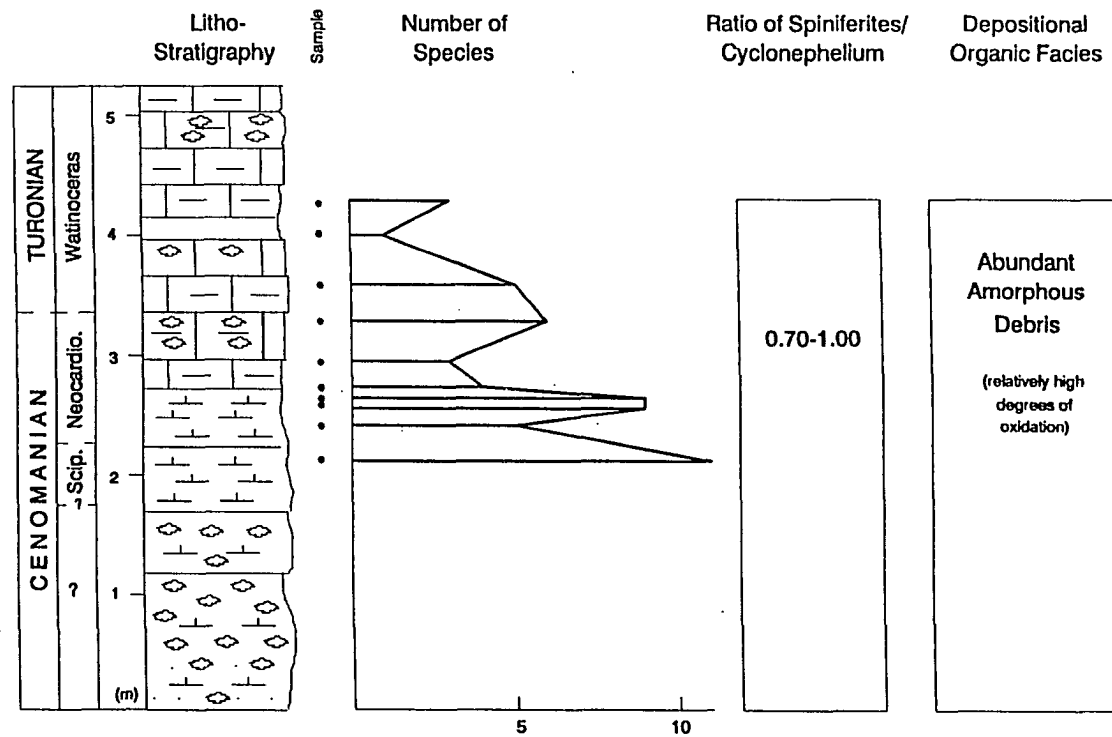


Fig. 24 Distribution of dinocyst species diversity, S/C ratio, and depositional organic facies in Ponca, Nebraska (N1).

Environmental Interpretation

Dinoflagellate assemblages of this section (Fig. 24) are much less diverse than those found at other sections, including the section with a similar rock association (C2). The generally abundant amorphous debris and very low percentage of terrigenous organic detritus throughout the sequence reflect a long-time stable continental shelf condition. The base of the section is characterized by calcareous shales with relatively high silt content, which was considered to be pre-*Sciponoceras* zone time (*Metoicoceras mosbyense* zone) based on the correlation of stable isotope stratigraphy and marker-beds (Pratt and Threlkeld 1984; Elder 1987b). The early period of marine transgression is suggested by the highest number of dinocyst species at the base of the *Sciponoceras* zone. The continued transgression (through the entire *Neocardioceras* and *Watinoceras* zones) is indicated not only by amorphous fecal debris facies, but also by increased bioturbation and increased carbonate productivity (Elder 1987b). Species diversity in this sequence shows a similar trend to those in other sections that decreases systematically from the basal transgressive interval to subsequent highstand interval, although the overall species richness is much lower. Unusually High organic carbon content (1.8-5.4%) in this section has been suggested mainly to be the result of low benthic oxygen concentrations based on depauperate benthic faunas (Elder 1987b). However, Elder also thought this

explanation as somewhat speculative, since high carbonate productivity itself, and bioturbation suggest a nearly normal benthic condition. High primary productivity due to upwelling seems to be another apparent interpretation for high organic carbon values. However, the lower degrees of dilution by the deposition of terrigenous detritus due to the position farther from shoreline should not be ruled out as a possibly factor.

(iv) Transitional Lithofacies Region Summary

Among three sections studied in this region, only the NM2 section is considered to represent the transitional lithofacies region, both lithologically and palynologically. The U2 section is better included in the clay-dominated lithofacies region due to its high silt content and low carbonate content, and resemblance of dinoflagellate assemblages to the U1 section; the N1 section looks more like a section in the carbonated-dominated lithofacies region (e.g., the C2 section) because of its higher carbonate content and poorly fossiliferous dinoflagellate assemblages.

The dinoflagellate assemblages in this region (NM2 section) show little difference in composition from the clay-dominated lithofacies region, except slightly lower diverse. The differences from the clay-dominated lithofacies region include (1) less cyclicity of palynologic parameters, particularly species diversity, which shows less response to cyclic marine transgression and regression, and (2) more

abundant *Spiniferites*-type cysts in the equivalent horizons which is represented by relatively higher S/C ratios, especially at the base of the *Sciponoceras* and *Neocardioceras* zones (S/C > 1).

The first transgression at the base of the *Sciponoceras* zone is suggested by an increase in the number of species and in S/C ratio within the matrix of abundant amorphous debris. The relatively smaller amplitude in this increasing trend and its association with a well-indurated marlstone bed rather than the basal transgressive lag sands indicate that rapid transgression occurred earlier in the region. Both the number of species and S/C ratio increase gradually and reach their highest in the middle or upper part of the *Sciponoceras* zone. Compared with the clay-dominated lithofacies region, the transitional lithofacies region shows a much less pronounced influence of the early transgression on dinoflagellate diversity. The number of species does not change significantly adjacent the *Sciponoceras*-*Neocardioceras* zonal boundary. The only indication of a possible regression at that time is a decrease in S/C ratio and a slight increase in inertinite.

The second transgression at the basal *Neocardioceras* zone is implied only by a significant increase in S/C ratio and the return of a pure amorphous debris facies. The moderately to highly diverse assemblages through the *Neocardioceras* zone indicate that the relatively long distance from the shoreline insulated this section from the strong influence of the

smaller fluctuations of relative sea level. As indicated in the clay-dominated lithofacies region, this region also does not show any indication of dinoflagellate extinction adjacent the *Sciponoceras-Neocardioceras* zonal boundary and the C-T boundary.

The continued transgressive interval and highstand interval through the upper *Neocardioceras* zone and the *Watinoceras* zone is indicated by the condensation of strata, increased abundance of macrofaunas, and a generally decreasing trend in dinocyst species diversity. The similar change in dinocyst diversity as in the clay-dominated lithofacies region, and in the coastal K-T section in Braggs (Habib et al., 1992) reflects the result of the change in depositional environment towards more inshore shelf or coastal conditions due to shoreline progradation during the highstand interval. The increase in current activity related to maximum transgression may have exerted additional influence on the diversity of dinocysts in the middle part of the *Watinoceras* zone.

In conclusion, the first transgression is well represented by an increase in both species diversity and S/C ratio. The second transgressive interval is only inferred by the S/C data, and to a lesser extent, by the change in depositional organic facies. The more abundance of the *Spiniferites* group than the *Cyclonephelium* group (S/C ratio >1) at the base of the *Sciponoceras* and *Neocardioceras* zone is

a good palynologic indicator of more offshore conditions during marine transgression. The lower concentration of landplant materials than at the clay-dominated sections even at the *Sciponoceras-Neocardioceras* zonal boundary suggest a limited influence of freshwater runoff from the western highland.

3. Carbonate-dominated Lithofacies Region

The sections studied in this lithofacies region are situated in the central seaway. Two sections, C2 (Pueblo) and C4 (Boulder), lie at the edge of the Denver Basin, approximately 700 km from both the eastern and western shorelines.

(i) Pueblo, Colorado (C2)

The section at Pueblo (C2) lies in the basal Bridge Creek Limestone Member of the Greenhorn Limestone (Elder and Kirkland 1985; Elder 1987b). This interval is part of the principal Cretaceous reference sequence of the Greenhorn Limestone, and consequently has received much attention. Some of the important and recent works which have dealt with Cenomanian-Turonian boundary event include Hattin (1971), Cobban and Scott (1972), Eicher (1977), Kauffman (1977b), Pratt (1981), Pratt and Threlkeld (1984), Cobban (1985), Hattin (1985), Leckie (1985), Pratt (1985), Watkins (1985), Zelt (1985), and Elder (1985, 1987b, 1989).

Lithology

The C2 section is characterized by cyclic alternations of calcareous or marly shale beds and marlstone or limestone beds. All 11 limestone beds (LS1-LS11) in the studied interval are well developed. Strata of the *Sciponoceras* zone contain well-laminated, well-indurated, gray calcareous or marly shales and four well-indurated gray limestone beds (LS1-LS4). Organic carbon contents in these shales are relatively low (0.3-1.8%). The increase in silt content at the base of the *Neocardioceras* zone which can be traced over a large part of the seaway may reflect the increased current activity or abundance of silt-sized foraminifera (Elder 1987b). The *Neocardioceras* zone is characterized by three well-indurated limestone beds (LS5-LS7) alternated with well-laminated, well-indurated, silty, marly and calcareous shales with moderate organic carbon content (1.5-2.0%) in the lower part, and with well-laminated, well-indurated, calcareous and marly shales with high organic carbon content (>4%) in the upper part. This increase in organic carbon content associated with a decrease in bioturbation in the upper *Neocardioceras* zone suggests decreased benthic oxygen concentration, which might be the main reason for faunal deterioration near the C-T boundary (Elder 1987b). The decrease in organic carbon content (<1%) and an increase in bioturbation at the basal *Watinoceras* zone suggests that increased benthic oxygenation was responsible for a faunal recovery and an influx of diverse Turonian

macrofaunas (Elder 1987b). Limestone bed in the lower *Watinoceras* zone (LS8) is typically massive, very well-indurated. The interval through the middle and upper part of the *Watinoceras* zone (LS9-LS11) is somewhat condensed based on the spacing of the limestone beds (Elder 1993, pers.comm.). The interbedded calcareous and marly shales are well-laminated and well-indurated. Organic carbon content in these shales increases from less than 0.6% at the base of the zone to between 2.0 to 2.5% in the upper part.

Fourteen samples were collected, including three from the *Sciponoceras* zone (2-meter-thick), five from the *Neocardioceras* zone (2.2-meter-thick), and six from the *Watinoceras* zone (2.1-meter-thick). All samples are from calcareous or marly shales except one from LS5 (Sample 18). The difference in carbonate content between this bed and the surrounding shales is small.

Dinoflagellate Stratigraphy

I. *Sciponoceras* Zone

Twenty-nine species and subspecies of dinocysts were recorded, only ten of them were restricted in this interval (Fig. 25). Amongst the abundant types are *C.distinctum*, *C.membraniphorum*, *S.ramosus ramosus*, *S.multibrevis*, *P.infusorioides*, *S.deformane* and *S.foliacea*. Relatively short-ranging species, such as *L.siphoniphorum* and

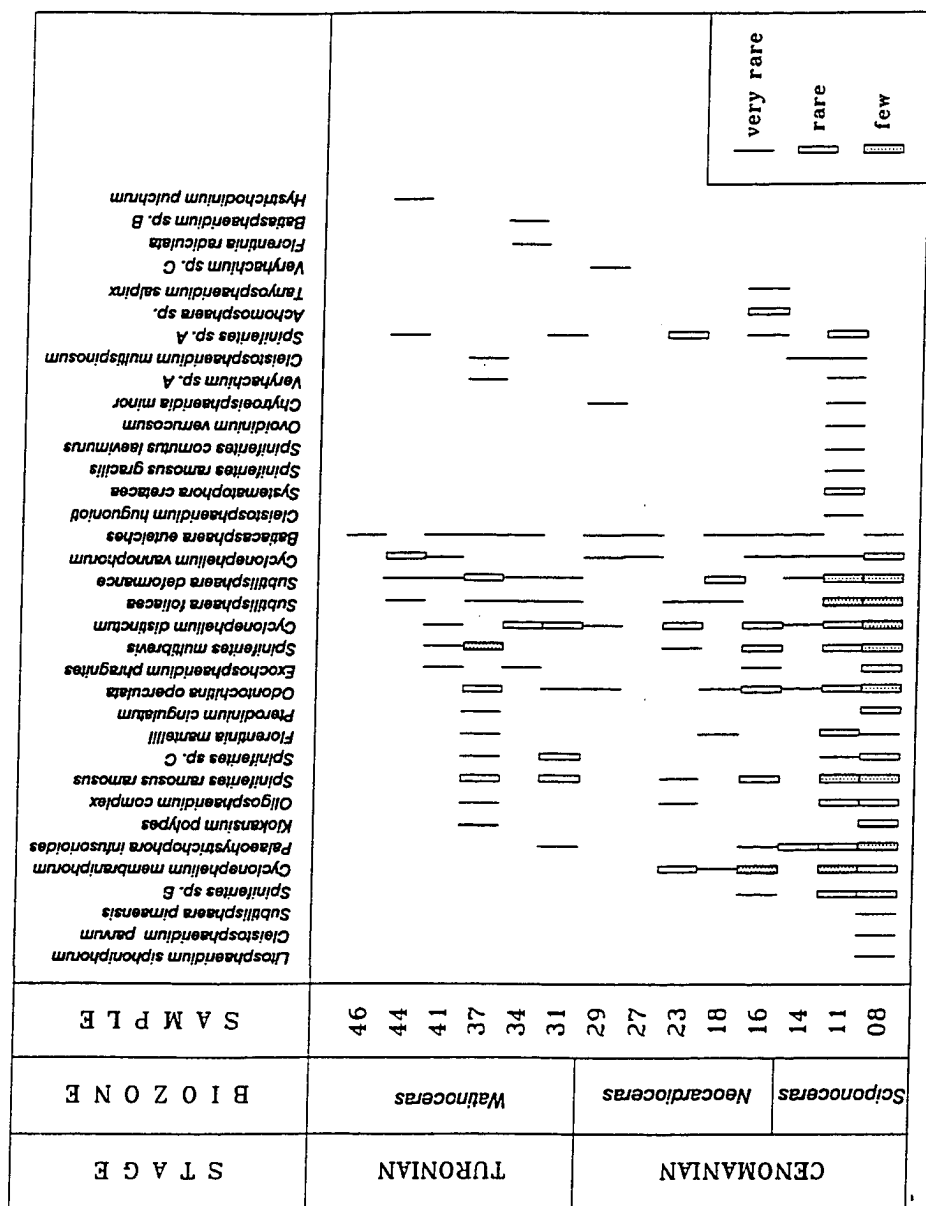


Fig. 25 Dinoflagellate-acritarch stratigraphy, Pueblo, Colorado (C2).

C. parvum, have their range tops within the *Sciponoceras* zone. Most species disappear temporarily in the upper part of the zone, and reappear in the overlying *Neocardioceras* zone.

II. *Neocardioceras* Zone

This zone yielded 19 taxa of dinocysts (Fig. 25). Sixteen were reported in the underlying *Sciponoceras* zone. The common species include *C. membraniphorum*, *C. distinctum*, *O. operculata*, *S. ramosus ramosus* and other relatively long-process forms, such as species of *Achomosphaera* and *Spiniferites*. Stratigraphically useful species *Chytroeisphaeridia minor* last appears in the upper part of the zone. A relatively diverse assemblage only occurs at the basal *Neocardioceras* zone. The rest of the zone is characterized by low species diversity and specimen abundance.

III. *Watinoceras* Zone

Among 21 taxa reported in this zone (Fig. 25), the most common are *S. multibrevis*, *O. operculata*, *S. deformance*, *C. vannophorum* and *S. ramosus ramosus*. Only three types first appear in this zone. Assemblage with more than ten species is only found in the middle part of the zone. The other parts are characterized by the poorly fossiliferous associations.

IV. Discussion

Most of the species recorded in this section are widely-

distributed forms which are also found in the clay-dominated and transitional lithofacies regions. The abundant and common types are still *Cyclonephelium* spp., *Spiniferites* spp. *Subtilisphaera* spp. and *Palaeohystrichophora* spp.

No single species, even the cyst type, dominates any assemblage. Unlike other sections (A1 and NM2), the assemblages in the *Sciponoceras* zone contain almost equally abundant *Cyclonephelium*-type, *Spiniferites*-type and peridinioid cysts. The specimen abundance for all three types decreases upsection, although not in the same degrees.

Both the last occurrence of *L.siphoniphorum* and *C.parvum* in the *Sciponoceras* zone, and the last occurrence of *C.minor* just prior to the *Neocardioceras*-*Watinoceras* zonal boundary suggest an age not younger than earliest Turonian.

Compared with clay-dominated or transitional lithofacies regions, this section yielded much lower diversity dinoflagellate assemblages. The species diversity reaches a maximum (20-22) in the lower part of the *Sciponoceras* zone, then decreases abruptly just prior to the *Sciponoceras*-*Neocardioceras* zonal boundary. Two thirds of species disappear temporarily at this interval, and reappear at the following *Neocardioceras* zone where species richness is back to a relatively high level.

Depositional Organic Matter

The section consists almost entirely of amorphous debris

(Fig. 26). The land-plant materials, including pollen and spores and vascular tissues, are insignificant. Dinoflagellate assemblages are either moderately diverse and abundant, or of low diversity and abundance. The approximate 10-20% (even higher in the *Neocardioceras* zone) amorphous debris is dark in color, which suggests relatively high degradation and oxidation. The increase in the amount of oxidized amorphous debris seems to be related to the decrease in species diversity (10-20 to less than 10 species).

Environmental Interpretation

The distribution of dinoflagellate and associated S/C ratio shows some cycles of change in depositional environments (Fig. 26). The first cycle occurs in the *Sciponoceras* zone. The increase in species richness, associated with abundant amorphous debris at the lower part of the zone, suggests an early period of marine transgression. Highly diverse and mixed macrofaunas with high species diversity represent the best biotic conditions and reflect sea level rise and transgression (Elder 1987b). The number of species and S/C ratio decrease adjacent the *Sciponoceras-Neocardioceras* zonal boundary, suggesting an environmental change possibly towards nearer-shore condition. The increase in silt content, a decrease in salinity (a negative shift in $\delta^{18}\text{O}$) and in benthic foraminiferal diversity in same interval were interpreted to be partly the result of increased current activity or

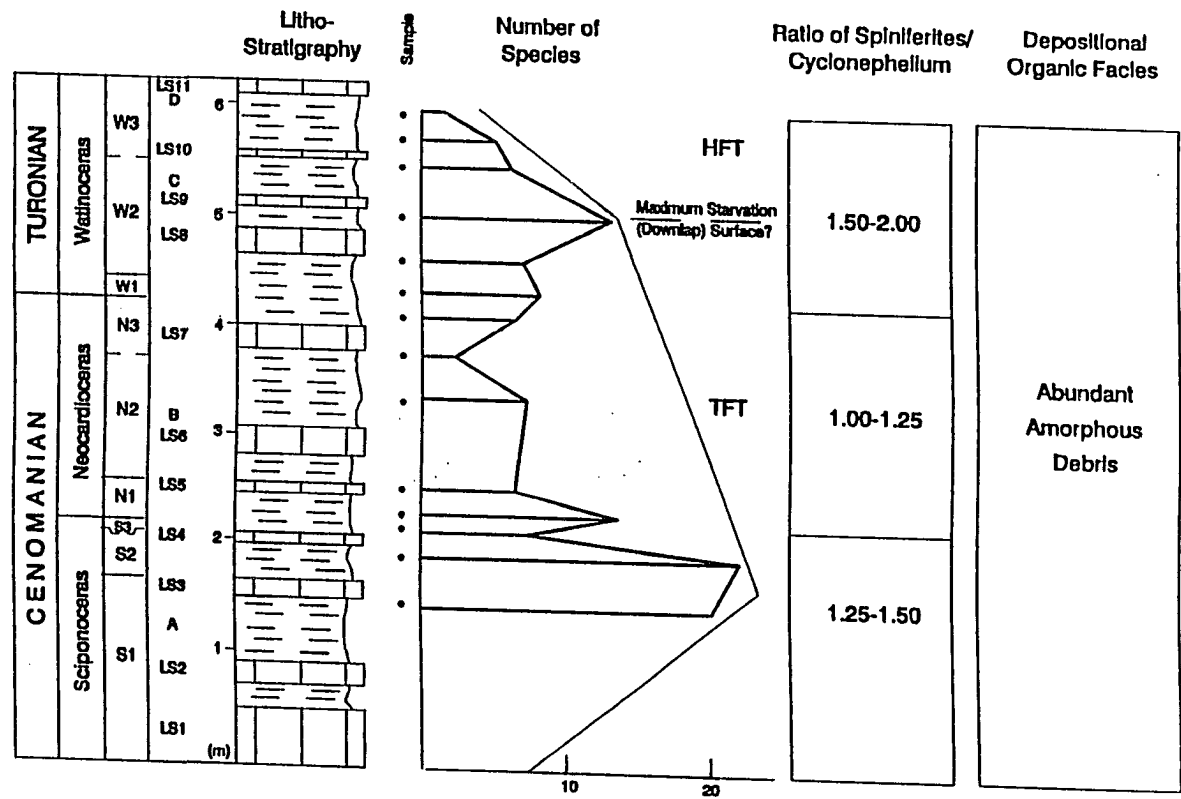


Fig. 26 Distribution of dinocyst species diversity, S/C ratio, and depositional organic facies in Pueblo, Colorado (C2).

abundance of silt-size foraminifera (Elder 1987b). Both a smaller increase in dinocyst species diversity and the recovery of a diverse inoceramid fauna (Elder 1987) occur simultaneously at the basal *Neocardioceras* zone. The low-diversity assemblages with relatively low S/C ratio through the rest of the zone, especially in the upper part of the zone, indicate a relatively restricted condition which might be in response to the fluctuations in salinity, oxygen and circulation pattern. The second apparent increase in species richness and high S/C ratios at the base of the *Watinoceras* zone may represent a continued transgression. The uncommonly high percentages of peridinioid cysts in some intervals may be related to increased current activity since these cysts can be carried longer time by currents than other cysts due to their air-bearing cavate morphology.

In conclusion, the species diversity and S/C ratio in this section show a similar pattern to those in the clay-dominated and transitional lithofacies regions, in spite of different diversity level. In the early transgressive interval, the number of species increases during the lower part of the *Sciponoceras* zone, reaches to the highest level at the middle part of the *Sciponoceras* zone, then decreases gradually through the late transgressive interval (upper part of the *Sciponoceras* zone and *Neocardioceras* zone) and the highstand interval (*Watinoceras* zone). The inflection point of the trend of species diversity occurs in the middle part of

the *Watinoceras* zone (around LS9), which was considered to be just above the base of the condensed interval (Elder 1993, pers. comm.). This is different from the nearshore A1 and the offshore NM2 sections, where the condensed interval occurs in the upper *Neocardioceras* zone (upper part of the condensed interval).

(ii) Boulder, Colorado (C4)

Section C4 lies in the Bridge Creek Limestone Member of the Greenhorn Limestone in the Denver Basin. Recent investigation includes previously measured four-meter-thick C-T boundary interval and newly measured portion (20 meter above and 7 meter below the previously measured portion). Forty-seven samples were collected during the visit by author, including 17 in the previously measured part, 23 and seven from the overlying and underlying strata, respectively. Only 20 samples, most in the C-T boundary interval, were processed and analysed.

Lithology

The section is characterized by olive gray, olive black, to brown black, poorly to moderately laminated, calcareous shales with several thin-layered, dark gray to olive black, limestones. These rocks appear to contain a constant content of organic carbon (1%) and carbonate content (40%), although both values slightly increase from the latest Cenomanian to

the earliest Turonian.

Dinoflagellate Stratigraphy

Twenty samples from latest Cenomanian (undifferentiated biozone) and early Turonian were analysed, and all proved to be devoid of dinoflagellate cysts. The absence of discernable dinocysts may be explained by unfavorable lithology which is not good for perservation of dinocysts. However, it is difficult to explain the presence of dinocysts in similar rocks of the C2 section, unless other factors, e.g., early diagenesis, are also accounted. Recent study shows that this section has undergone some degrees of thermal alteration (Elder 1993, pers.comm.), which probably result in the loss of dinocysts in sediments.

Depositional Organic Matter

The section is characterized by a typical amorphous debris facies. Terrestrial organic materials, where found, are mostly degraded and fragmented, and constitute a very small part of the organic residues.

Environmental Interpretation

The relatively stable shelf condition through the C-T interval is suggested by abundant amorphous fecal pellets and low concentration of landplant materials within carbonate-rich deposits. Some of amorphous fecal debris (10-20%) has been

oxidized, suggesting a possibility that either surface or bottom water conditions may have not been always anoxic.

(iii) Carbonate-dominated Lithofacies Region Summary

Samples from the high-carbonate marly shales, marlstones and limestones in this carbonate-dominated lithofacies region display the least diverse and most sparse assemblages. The N1 section (eastern transitional lithofacies region) is summarized here due to the similarity of lithology and dinoflagellate assemblages to the C2 and C4 sections.

The cyclicity of palynologic criteria in response to marine transgression and regression is similar to that in the transitional lithofacies region, even though the amplitude is much smaller. The basinwide transgression at the base of the *Sciponoceras* zone is indicated by an increase in species diversity and abundant amorphous debris. The small decrease in the number of species and in S/C ratio near the *Sciponoceras-Neocardioceras* zonal boundary suggests a change in depositional environments possibly in response to a short period of stillstand of sea level. The generally low-diversity assemblages with relatively lower S/C ratios, especially in the middle to upper part of the *Neocardioceras* zone, reflect a relatively restricted environment which might be the result of increased terrigenous detrital influx. The relatively high-diversity assemblages with increased S/C ratio in the *Watinoceras* zone indicate a return of normal marine condition.

As recorded in the other two lithofacies regions, this region also shows the same decreasing trend in species diversity from the lower transgressive interval through the highstand interval, reflecting the distant shoreline progradation. The relatively higher concentration of oxidized amorphous debris in most samples indicates that the processes of oxidation and degradation have taken place either on the route descending from the surface water or on the bottom of the basin.

Summary of the Western Interior Basin

The preservation of dinoflagellate cysts is highly variable in the sections investigated and closely corresponds to lithology. The best preservation was observed in the calcareous and marly shales of the clay-dominated lithofacies region. In the carbonate-dominated lithofacies area, many samples proved to be devoid of dinocysts or the dinocysts were poorly preserved. In general, dinocysts are more diverse and numerous in the clay-dominated lithofacies region than in the carbonate-dominated lithofacies region. The transitional lithofacies region shows a transitional appearance between the clay-dominated and the carbonate-dominated regions in both species diversity and specimen abundance, but it is more like the former region, particularly in the western part of the basin.

Most species and subspecies recorded in the Western

Interior Basin are relatively long-ranging and widely distributed forms which have been previously reported in many European and other sections. The frequent appearance of some relatively cosmopolitan forms, such as *Odontochitina operculata*, *O.costata* and *Oligosphaeridium complex*, and the rare occurrence of domestic types in most samples reflect a good connection between the relatively restricted Western Interior Seaway and other marine basins during Cenomanian-Turonian marine transgression, particularly through the south. This is further supported by widespread, diverse macrofaunas migrated from southern warm-temperate to subtropical epicontinental seas.

The occurrence datums of several dinoflagellate species, some as shown in the standard biohorizons (Chen et al., 1988), can be used for dating and correlation. They include the last occurrences of *Litosphaeridium siphoniphorum*, *Trithyrodinium suspectum*, *Cleistosphaeridium parvum* and *Gingiodinium spinulosum* in the *Sciponoceras* zone, and the last occurrences of *Surculosphaeridium longifurcatum*, *Ascodinium acrophorum*, *Achomosphaera sagena* and *Chytroeisphaeridia minor* in the *Neocardioceras* zone. These indicate an age not later than latest Cenomanian. The last occurrences of *Epelidosphaeridia spinosa* and *Scrinioidinium campanula* at the base of the *Watinoceras* zone suggest an early Turonian age.

The most common species are *Cyclonephelium distinctum*, *C.vannophorum*, *Odontochitina operculata*, *Spiniferites ramosus*

ramosus, *Exochosphaeridium phragmites* and *Palaeohystrichophora infusorioides*. They occur in most samples and predominate in different intervals in different sections. The other common types include *Florentinia mantellii*, *Oligosphaeridium complex*, *Spiniferites multibrevis*, *Subtilisphaera deformance*, *S.foliacea*, *Kiokansium polypes*, *Pervosphaeridium pseudohystrichodinium*, *Odontochitina costata*, and *Cyclonephelium membraniphorum*.

The trends in species diversity and numerical abundance show a number of significant features. Some of these differences are explained by the lithological variation. Samples from sediments with high carbonate content, for example, invariably contain poor assemblages or are even barren (5 to 10 species or even less), whereas most samples from sediments with low carbonate content contain much richer assemblages (20 to 30 species or even more). Dinocysts are numerous in most of the samples from the clay-dominated and transitional lithofacies regions but are scarce in most of those from the carbonate-dominated lithofacies region. In the carbonate-dominated lithofacies region, the numerical abundance of dinocysts changes with a fluctuation of carbonate content. The factors controlling the concentration of dinocysts in sediments are complicated. Carbonate content in sediments is believed to be a major possibility. The correlation of cyst concentration and carbonate content will be discussed in the later chapter.

Detailed study reveals that the number of species is mostly affected by the change in depositional environments resulting from marine transgression and regression. The significant increase in species diversity represents a rapid expansion of the continental shelf during marine transgression, whereas a sharp decrease in the number of species reflects a regressive pulse during the period of stillstand or drop of sea level. In the Western Interior basin, the curves of species diversity in most sections show two major peaks in the sequence. The first, occurring at the base of the *Sciponoceras* zone, is prominent and developed basin-wide. The apparent increase in species richness associated with abundant amorphous debris indicates a rapid transgression and the development of extensive epicontinental seas. The subsequent decrease in species diversity is associated with increased concentration of landplant materials adjacent the *Sciponoceras-Neocardioceras* zonal boundary. It is only well-developed in the nearer-shore areas, particularly along the western margin. This suggests that only the western part of the basin (nearer-shore) was influenced by the regression. The second marine transgression is identified by multiple occurrence datums immediately above poorer assemblages, and a pure amorphous debris facies, in the basal *Neocardioceras* zone. This event reflects a relatively smaller-scale transgressive pulse. The second decrease in species diversity adjacent the C-T boundary is even smaller than the

first one in the nearshore sections, and suggests a diminished effect of regression. The magnitude of change in species diversity reflects, to some extent, the range and intensity of marine transgression and regression. The first transgression during a rapid sea level rise is developed after a period of a relatively lowstand interval, and therefore, the influence on dinocyst diversity is extensive through the whole basin. The second transgression is developed after a stillstand period within a generally increasing transgression, and therefore, the effect on dinocyst richness is locally-restricted and less-intensified.

Dinoflagellate cysts (in all sections studied) do not show any level of mass extinction. However, they do show a decrease in species diversity adjacent the *Sciponoceras-Neocardioceras* zonal boundary and the C-T boundary. Most species disappear temporarily near these boundaries, and then reappear immediately above. The decrease in species richness is in response to environmental change, possibly related to the stillstand or fall of relative sea level and to increased shoreline progradation.

Species dominance is generally low. No single or few species constitute more than 25% of any assemblage. Nowhere in the basin is there an assemblage that is overwhelmingly dominated by a single species. This suggests that depositional environments are more likely open marine even in nearer-shore areas.

Three morphotype groups are defined based on the similarity of morphology and ecological significance of the previously widely-used dinoflagellate species (Fig. 27). The *Spiniferites* group includes several long, complex-process chorate cysts with major components of *S. ramosus ramosus*, *S. multibrevis*, *A. ramulifera*, *Hystriosphæridium* spp., *O. complex*, and other species of *Spiniferites*. They are considered to be characteristic of a more offshore, normal marine condition. The *Cyclonephelium* group includes some short, simple-process proximate forms with major components of *C. distinctum*, *C. vannophorum*, *E. phragmites*, *K. polypes*, and species of *Cleistosphæridium* and *Sentusidinium*. The abundance of this group represents a more near-shore, less salinity condition. The peridinioid group is composed of some cavate cysts, such as *P. infusorioides*, and species of *Subtilisphaera* and *Isabellidinium*. The abundance of such more euryhaline and upwelling-related peridinioid cysts increases towards nearshore conditions. Quantitatively, these three groups jointly constitute 60-80% or even more of all dinoflagellate cysts. Figure 28 shows the comparative study of the frequency change in *Spiniferites* and *Cyclonephelium* groups from the A1 and NM2 sections. These two groups are considered to be more useful and meaningful in the determination of depositional environments based on (1) an apparent reverse correlation between the abundance of *Cyclonephelium* group and of *Spiniferites* group. Where the number of one group increases,

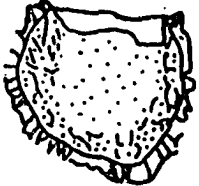
	MORPHOLOGY		COMPONENTS	ENVIRONMENTAL SIGNIFICANCE
Spiniferites Group		Chorate cysts with long and complex processes	<i>Spiniferites</i> spp. <i>Achomosphaera</i> spp. <i>Oligosphaeridium</i> spp. <i>Hystichosphaeridium</i> spp. <i>Hystichodinium</i> spp. <i>Avellodinium</i> spp.	Open marine shelf environments
Cyclonephellium Group		Proximate cysts with short, stout and barbed processes	<i>Cyclonephellium</i> spp. <i>Exochosphaeridium</i> spp. <i>Cleistosphaeridium</i> spp. <i>Kiokanslum</i> spp. <i>Michystridium</i> spp. <i>Sentusidinium</i> spp.	Costal to nearer-shore relatively restricted environments
Peridinioid Group		Cavate cysts with smooth wall or low-relief ornamentations	<i>Palaeohystrichophora</i> spp. <i>Subtilisphaera</i> spp. <i>Palaeoperidinium</i> spp. <i>Luxadinium</i> spp. <i>Gingnodinium</i> spp. <i>Gelselodinium</i> spp.	Euryhaline forms indicate estuarine and lagoonal environments

Fig. 27 The morphology and composition of the Spiniferites, Cyclonephellium and peridinioid groups, and their environmental significance.

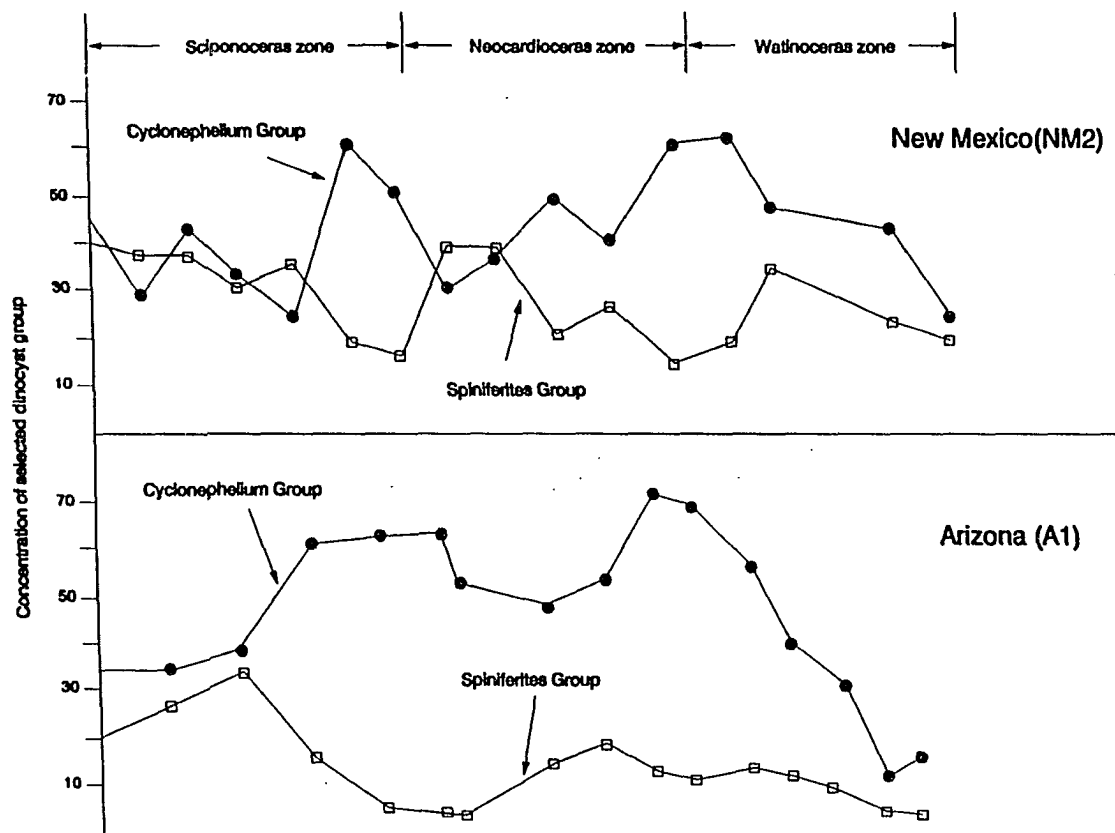


Fig. 28 Comparison of two selected dinocyst groups between the A1 and NM2 sections.

it is always accompanied by the decrease of another group; and (2) the relative frequency of *Spiniferites* group is always higher in the NM2 section than in the A1 section, suggesting an increasing trend towards basin center. Therefore, the ratio of *Spiniferites/Cyclonephelium* (S/C ratio), based on the relative abundance of each group, is a more quantitative measurement than any specific cyst or cyst-group.

A summary of the relative frequencies of three selected cyst groups in the three biozones is shown in Table 3 and Figures 29-31. The *Cyclonephelium* group is the most abundant in most samples, except in some intervals of the carbonate-dominated lithofacies region (C2). Its numerical abundance in assemblages fluctuates with varying *Spiniferites* group, and decreases in offshore direction. The average percentage for this group (see Table 3) in an assemblage decreases from 49 in the nearshore A1 and U1 sections, through 41 in the offshore NM2 section, to 29 in the basin-center C2 section. In contrast, the *Spiniferites* group increases its concentration in offshore direction from average 20 percent in the A1 and U1 section, to 30 percent in the NM2 section, and to almost 40 percent in the C2 section. The peaks of relative frequency for *Cyclonephelium* group are accompanied with the lowest values of *Spiniferites* near the *Sciponoceras-Neocardioceras* zonal boundary, and to a lesser extent, the Cenomanian-Turonian boundary. This reflects the increased influence of terrestrial materials during possible marine regressions or shoreline

SECTION	ARIZONA (A1)			UTAH (U1)			NEW MEXICO (NM2)			COLORADO (C2)		
	S	N	W	S	N	W	S	N	W	S	N	W
Spiniferites Group (%)	22	15	13	28	19	22	32	31	26	35	42	35
Cyclonephelium Group (%)	41	58	49	49	47	48	39	42	43	27	40	20
Peridinioid Group (%)	2	8	9	9	21	15	5	12	18	25	6	24
RATIO of S/C	0.53	0.26	0.27	0.57	0.41	0.48	0.82	0.74	0.60	1.30	1.05	1.75
RATIO of S/(C+P)	0.51	0.23	0.22	0.48	0.28	0.35	0.73	0.57	0.43	0.67	0.91	0.80

Table 3 Average percentage for selected dinocyst groups, ratio of Spiniferites to Cyclonephelium and ratio of Spiniferites to combined Cyclonephelium and peridinioid from the four sections studied.

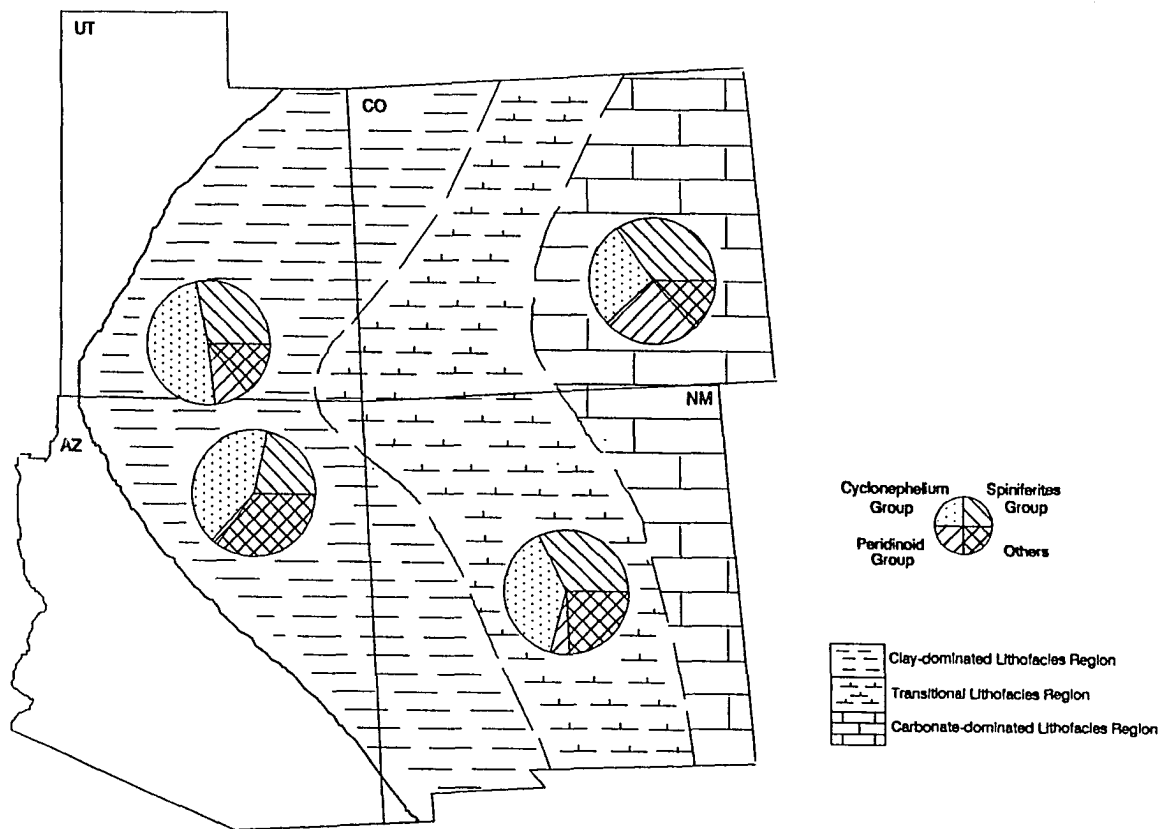


Fig. 29 Ratios of selected dinocyst groups in the Sciponoceras zone from the U1, A1, NM2 and C2 sections.

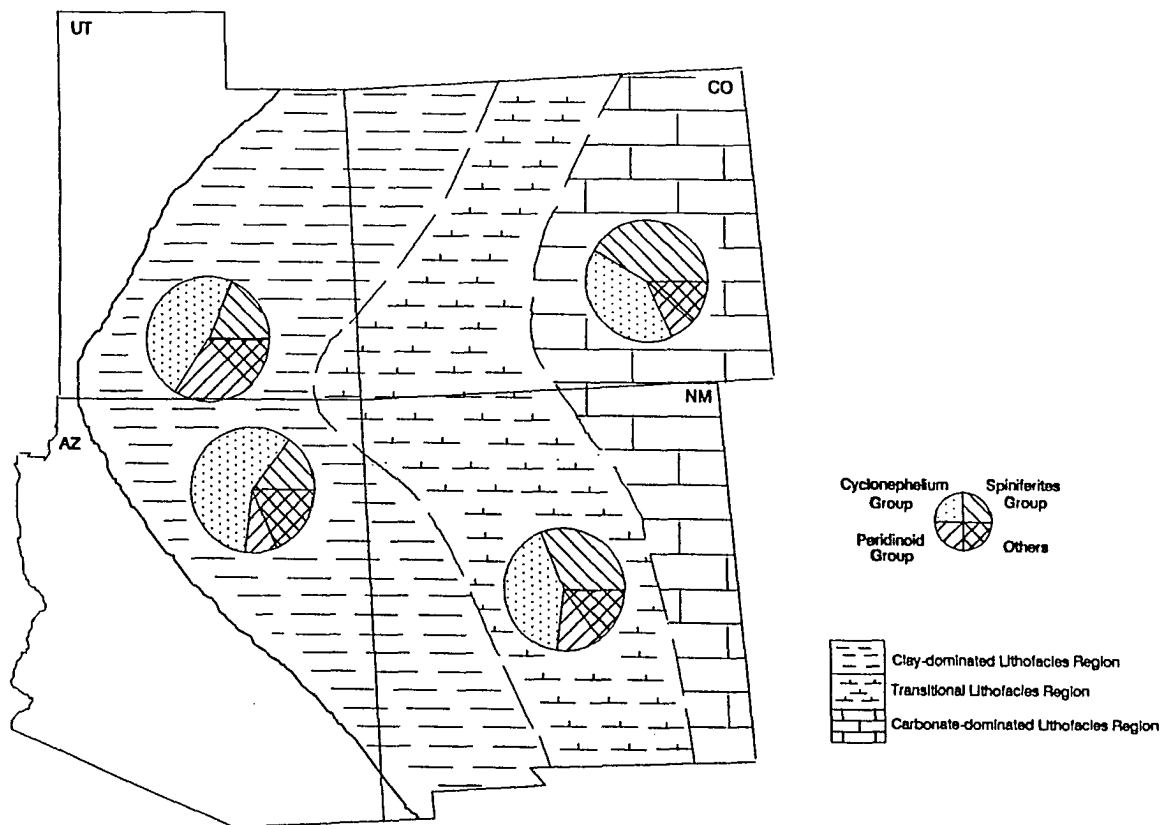


Fig. 30 Ratios of selected dinocyst groups in the Neocardioceras zone from the U1, A1, NM2 and C2 sections.

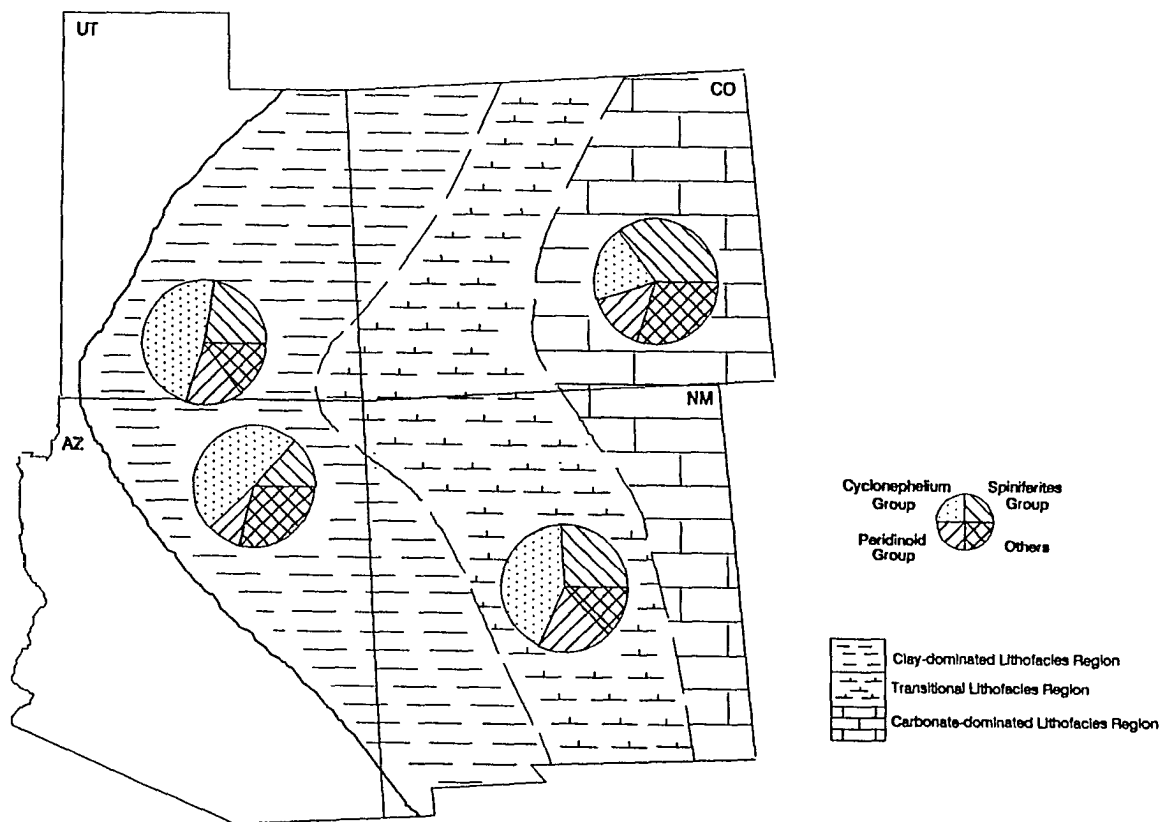


Fig. 31 Ratios of selected dinocyst groups in the Watinoceras zone from the U1, A1, NM2 and C2 sections.

progradation. The significant increases in the abundance of *Spiniferites* group occur at the base of biozones, especially at the beginning of the *Sciponoceras* and the *Neocardioceras* zones. This is a good palynologic response to marine transgressions during the rise of sea level (Li and Habib 1993).

The curves of the S/C ratio from different lithofacies regions show a more accurate diagram in terms of the changes in depositional environments caused by marine transgression and regression. The highest S/C value occurs at the basal *Sciponoceras* zone in all above four sections (Table 3), indicating sea level rise and a rapid transgression in the whole basin in early *Sciponoceras* zone time which is also evidenced by the transgressive lag sands in nearshore areas and a sudden increase in faunal component from the south. The base of the *Neocardioceras* zone is also characterized by an increase in S/C ratio, but the amplitude of increase is relatively smaller compared with the first one. This suggests that the second transgressive pulse in early *Neocardioceras* zone time is only of a smaller-scale, and not well-developed in the whole area. The S/C ratio also shows an increase in the seaward direction. Figure 32 shows a comparison of S/C ratios in the equivalent horizons between the A1 (nearer-shore) and the NM2 (off-shore) sections. The curves of S/C ratio from both sections look very similar. That the value in the NM2 is always higher than the comparable part in the A1 suggests that

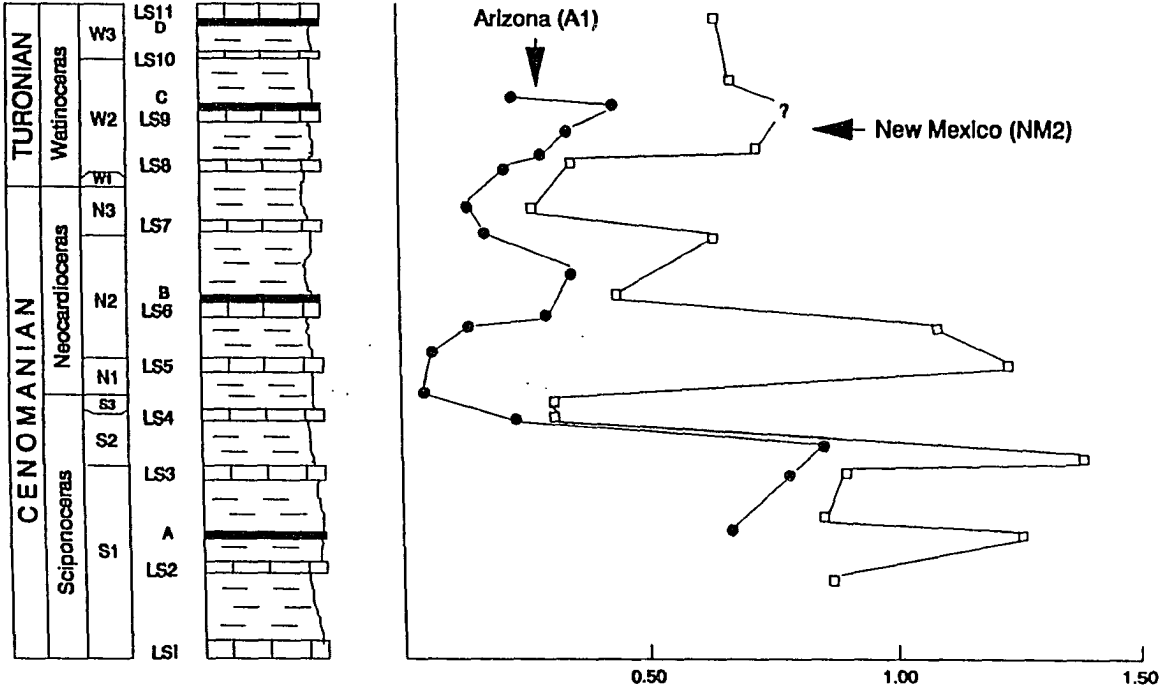


Fig. 32 Comparison of the S/C ratio in the equivalent horizons between clay-dominated lithofacies A1 and transitional lithofacies NM2 sections.

S/C ratio is a very good indicator of environmental changes from nearshore to offshore. Both sections exhibit medium values at the base of the sequence gradually increasing in the lower and middle parts of the *Sciponoceras* zone which represents increased effect of marine transgression and deepening of the water. The S/C ratio decreases suddenly in the upper part of the *Sciponoceras* zone, and reaches the lowest in the interval near the *Sciponoceras-Neocardioceras* zonal boundary. This trend combined with relatively high concentrations of land-plant materials suggests a nearer-shore environment. The return of S/C values to a relatively high level at the beginning of the *Neocardioceras* zone indicates a new round of transgression. The disproportionally lower S/C values in the A1 section during this interval may not be explained only by sea level change. Increased sedimentation rates and decreased salinity in surface waters resulted from uncommonly large amount of freshwater influx might have led to an disproportional increase in the abundance of *Cyclonephelium*-type cysts. The lower S/C ratio is also found at the Cenomanian-Turonian boundary. It is different from the previous *Sciponoceras-Neocardioceras* zonal boundary in that the decrease in the S/C ratio is not accompanied by a concurrent decrease in species diversity, especially in the NM2 section. This suggests that the S/C ratio can reflect a more detailed, smaller-scale change of depositional environments that species diversity may not indicate. A

similar cycle observed in the *Watinoceras* zone with a distinct increase followed by a slight drop in the S/C ratio suggests an oscillation of depositional environments from onshore into a more offshore direction to return of nearer-shore.

The application of the peridinioid cysts in the environmental analysis is somewhat controversial based on the previous studies. It has been argued that the abundance of peridinioid cysts in assemblages necessarily reflects a more nearer-shore environment (e.g., Brinkhuis and Zachariasse 1988). Detailed study from above four sections (Table 3) shows that the peridinioid group is only a minor component in most dinocyst assemblages except in the C2 section where this group is more common, especially in the *Sciponoceras* and *Watinoceras* zone. The high concentration of the peridinioid cysts may reflect a stronger influence of increased current activity than environmental change towards nearshore. Therefore, the interpretation based only on the abundance of the peridinioid cysts is somewhat dangerous. In this study, the peridinioid and the *Cyclonephelium*-type cysts are treated as a collective group in the environmental analysis. The ratio of S/(C+P) has the same environmental indication as the ratio of S/C. That means that the higher the value, the more open marine condition. The S/(C+P) values from above four sections (Table 3) show a very similar trend (decreasing from onshore to offshore) to the S/C ratios. Only one exception is found in the *Sciponoceras* zone, where the NM2 section has a higher

average ratio than the C2 section. However, this ratio proves to be useful, particularly when *Cyclonephelium*-type cysts are not well-developed.

The organic facies analysis for all samples collected in the Western Interior Seaway indicate that all major types of organic matter (palynomorphs, phytoclasts and amorphous particles) are found in the Cenomanian-Turonian sediments. Amorphous debris comprises the largest (80-90%) single type. In the clay-dominated region, dinoflagellate assemblages are relatively diverse and abundant. Species diversity fluctuates with the changes in concentration of amorphous debris and inertinite. In the basin center (carbonate-dominated lithofacies region), dinoflagellate assemblages are meager, and associated with a relatively unchanged organic facies. Inertinite is important in the intervals of the *Sciponoceras-Neocardioceras* zonal boundary and the CT boundary, particularly in the clay-dominated region. Most of the changes in dinoflagellate species diversity are correlated with the increase in the amount of inertinite.

In conclusion, the distribution of dinoflagellate assemblages in the whole basin shows some common features (Fig. 33). However, different lithofacies regions display several different dinoflagellate characteristics through the study interval. The S/C ratios show a clear trend from less than 1 in the nearer-shore clay-dominated region, fluctuating around 1 in the transitional lithofacies region, to always

	Clay-dominated Region	Transitional Region	Carbonate-dominated Region
Preservation	Well-preserved diverse and numerous	Well-preserved moderately-highly diverse	Poorly-preserved or devoid of dinocysts
Species Diversity	High fluctuates with marine trans- gression-regression cycles	Medium-High Less fluctuations	Low little change
Specimen Abundance	High	Medium-High	Low
	decreases with increasing carbonate content →		
Dominance by few species	Low	Low	Low
Dominance by cyst groups	Cyclonephelium group always outnumbers Spiniferites group	Cyclonephelium group more abundant than Spiniferites group except at the basal S. and N. zone	Number of Cyclonephelium smaller than Spiniferites group
S/C ratio	< 1 similar fluctuations with species diversity	< 1 - > 1 > 1 only at the base of the S. and N. zone	> 1
Extinction Event	← a drop in species diversity and abundance near S/N and CT boundaries	No extinction event throughout the sequence	→
Ratio of land/mar. OM	Low	Very low	very low
Depositional Organic Facies	Abundant amorphous debris, with inertinites near the S/N zonal and CT boundaries	Pure amorphous debris	Very pure amorphous debris

Fig. 33 Summary of palynologic data in the three lithofacies regions.

larger than 1 in the carbonate-dominated region. Species diversity in the transitional region is more like that in the clay-dominated region (more diverse), and is apparently different from the carbonate-dominated region. Both the transitional and carbonate-dominated regions show a similar composition of depositional organic matter, with overwhelmingly dominant amorphous debris, whereas in the clay-dominated region, the abundance of amorphous debris decreases with an increased concentration of inertinite. These differences are largely controlled by the distance away from the paleoshoreline during marine transgression and regression, particularly in the western part of the basin.

The lower part of the *Sciponoceras* zone (late Cenomanian) is characterized by a pronounced increase in species diversity, numerical abundance and S/C ratio within a matrix of amorphous debris. The presence of an increasing trend in species diversity is a good indicator of the development of the basal transgressive interval. This is supported by transgressive lag sands in the nearer-shore areas. The middle and upper parts of the zone are characterized by diverse and abundant assemblages, suggesting a continued transgression. The first significant decrease in species richness and specimen abundance is found at the top of the zone associated with a decrease in S/C ratio in the clay-dominated lithofacies region, suggesting the period of marine regression in response to a stillstand or drop of sea level. A similar cycle is also

found in transitional lithofacies region, although the decreasing trend is only suggested by a decrease in S/C ratio. The amount of terrigenous organic detritus increases at the *Sciponoceras-Neocardioceras* zonal boundary in the western clay-dominated lithofacies region. This reflects a higher fluvial discharge from the highland to the west which is indicated by a decrease in $\delta^{18}\text{O}$ values and in the illite-smectite mixed-layer clay ratio (Pratt 1985).

The second transgressive pulse at the base of the *Neocardioceras* zone (latest Cenomanian) is indicated not only by an onshore shift of the sediment depocenter and an influx of the diverse southern faunas, but also by a remarkable increase in dinoflagellate diversity. This transgression is palynologically considered to be locally developed and on a relatively smaller scale because transgression-bounded multiple occurrence datums above lowstand- or stillstand-bounded poorly fossiliferous assemblages only occur in the nearshore areas. It is followed by a decrease in species richness and S/C ratio adjacent the C-T boundary in the nearshore Arizona section. The absence of such a change in the nearshore U1 section suggests that the second regression, if it exists, should be more locally developed. The relatively high concentration of landplant particles through the entire *Neocardioceras* zone in the clay-dominated lithofacies region supports reduced surface water salinity as a result of large amounts of fresh-water influx from the western highland (e.g.,

Pratt 1985).

The early Turonian *Watinoceras* zone is characterized by the low to moderately diverse and abundant assemblages associated with widespread amorphous fecal debris in all lithofacies regions. This pattern may be correlated with the development of carbonate-rich sediments in the highstand interval.

DISCUSSION

The Cenomanian-Turonian Extinction Event

The Cenomanian-Turonian boundary marks the second-largest extinction event in the Cretaceous period. Detailed study has revealed that the event in the Western Interior basin is represented by a series of short-term, ecologically graded extinction steps (Kauffman 1984a, 1984b, 1988; Elder 1985, 1987a, 1987b, 1989). It has been thought to be associated with abrupt major marine paleobiogeographic changes, as well as environmental perturbations, such as a global oceanic anoxic event, rapid changes in temperature and in marine chemistry (Kauffman 1984a, 1984b). Detailed geochemical analyses reveal extraordinarily large stable isotope ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$) perturbations, as well as organic carbon and Mn fluctuations across the boundary. They are considered to be reflections of dynamic changes in temperature, salinity, marine carbon cycle, and Mn deposition (Pratt 1985; Zelt 1985; Arthur et al., 1987; Schlanger et al., 1987). Orth et al. (1988, 1993) conducted a high-resolution stratigraphic sampling and geochemical analysis, and found an Iridium enrichment zone containing two spikes below the C-T boundary. Each Ir spike correlates precisely or very closely with one of the first two steps of

the C-T mass extinction interval (Elder 1987a, 1987b; Kauffman 1984a, 1984b, 1986, 1988). Unusually high iridium and other trace-element abundances found in C-T strata in the Western Interior seaway can be explained either by extraterrestrial mechanisms (Orth et al., 1988), or by an earthbound source (Elder 1989; Orth et al., 1993).

Elder's data (1987a, 1987b, 1989) showed eight major, abrupt, molluscan extinction steps spanning the C-T boundary interval in the Western Interior basin. Six of these steps occur within the interval of this study. Approximately 13 percent of 61 genera and 51 percent of 81 species became extinct through this interval. Similar faunal turnovers are also recorded in other tropical and temperate areas (Kauffman 1984a; Hilbrecht 1986; Jarvis et al., 1988a). Unlike the benthonic-dominated molluscan fauna, planktonic foraminifera exhibit only one step of extinction (*Rotalipora* extinction). This approximately coincides with the first step of molluscan extinctions (Eicher and Diner 1985; Leckie 1985). The extinction of the deep-dwelling foraminifera *Rotalipora* has been interpreted to be a result of the oxygen minimum zone expanding upward into its depth range in the open ocean (Leckie 1985; Jarvis et al., 1988a; Hart and Leary 1989; Tocher et al., 1989).

In the Black Mesa Basin (Arizona), 93 percent of the 28 species of ammonites and inoceramids become extinct in the C-T boundary interval (Elder 1987a, 1987b). Study of calcareous

nannoplankton from the same material reveals that there is no clear biotic turnover of this group of nannofossils (Bralower 1988). Most of the marker species range right through the C-T boundary, and less than 5 percent either become extinct or first appear in this interval (Bralower 1988). The reason why the planktonic nannofossil assemblages do not suffer from the dramatic changes that beset the macrofossils was attributed to the limited effect of anoxia caused by either an expansion of a mid-water oxygen minimum zone or unusually high fresh water runoff (Bralower 1988). The extinction of the planktonic foraminifera *Rotalipora* has not been recorded in the Black Mesa Basin. *Rotalipora* is generally absent, which suggests the shallower mid-shelf environments of this region (Leckie et al., 1991; Kirkland 1991).

The fossil record of dinoflagellates does not provide significant extinctions, even in several important boundaries like those of the Permian-Triassic and Cretaceous-Tertiary (Bujak and Williams 1979). The change in dinoflagellate diversity, rather than extinction, is the more common case (Moshkovitz and Habib 1993). Factors which cause change of dinoflagellate diversity include sea-level change, nutrient depletion, climatic deterioration, cosmic radiation, compositional changes in sea water and biological interactions (Bujak and Williams 1979). Of these, sea level change, and to a lesser extent, temperature fluctuation are the most important controls (Tappan and Loeblich 1972; Bujak and

Williams 1979).

The dinoflagellate data, along with the previous macrofossil and nannofossils evidence, from the Blue Point section (Arizona) are plotted in Figures 34 and 35. The distribution pattern of dinocysts throughout the C-T boundary sequence seems different from both that of nannofossils and macrofossils. It differs from the latter in lack of a series of extinctions. Most important dinoflagellate species disappear temporarily prior to or at the *Sciponoceras-Neocardioceras* zonal boundary and the Cenomanian-Turonian stage boundary, then reappear in the above strata, respectively (Fig. 34). In contrast to the calcareous nannofossils, dinoflagellate assemblages show a dramatic change in species diversity and specimen abundance near both boundaries (Fig. 35). Elder (1987a, 1987b, 1989) thought that many of the molluscan extinctions reflected regional environmental deterioration patterns associated with the low benthic-oxygen and reduced salinity conditions during periods of clay-rich deposition. Elder (1987b, 1989) also considered the first and largest increase in macrofaunal turnover rates (extinction step 2) as being coincident with the freshening in the upper water column (a major negative excursion of stable oxygen isotope ratios). The second abrupt faunal replacement (extinction step 5) coincides with the return of $\delta^{18}\text{O}$ values to those more characteristic of normal marine conditions. Kauffman(1988) has interpreted these two events as the start

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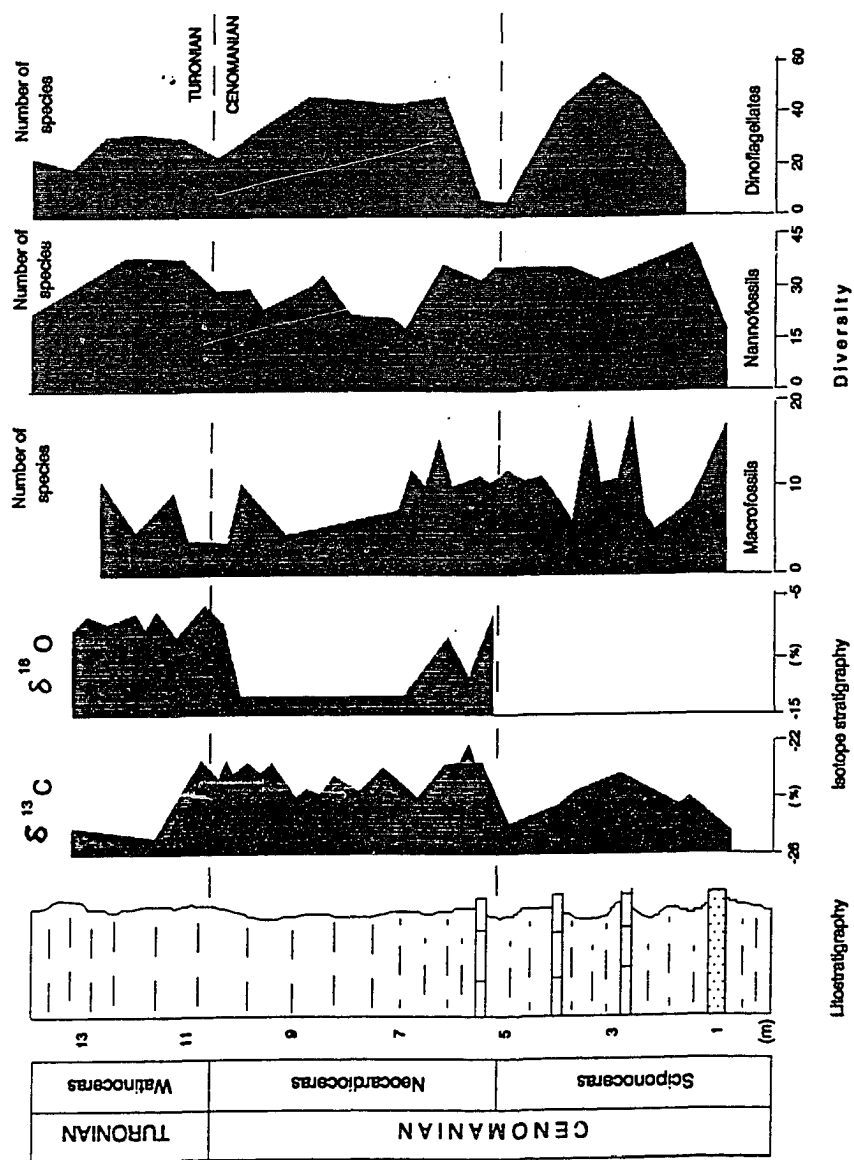


Fig. 35 The lithostratigraphy, isotope stratigraphy and diversity of macrofossils and microfossils in the Black Mesa, Arizona. Lithostratigraphic data from Elder 1987b, 1989; isotope data from Pratt 1985.

Calcareous nannofossil data from Bralower 1988; macrofossil and

and the end of the North American Desalination Event. The studies from the European sections show that these two extinction steps (bottom and top of the *Neocardioceras* zone) might be of global scale (Elder 1989). The remaining steps (step 1, 3, 4, and 6) apparently reflect cyclic changes in water mass and substrate characters in the Western Interior basin produced in response to orbital forcing of climate (Kauffman 1988; Elder 1989). The reduction of oxygen level on the seafloor during the *Neocardioceras* zone time could be a result of either the establishment of a brackish water cap, or an excursion of the oxygen minimum zone into the seaway, or both (Kauffman 1984b; Elder 1987b; Leckie et al., 1991). The oxygen minimum zone might be restricted to the deeper central part of the Western Interior seaway because of the existence of a moderately diverse benthic invertebrate and foraminiferal fauna in the Black Mesa basin (Kirkland 1990). The variation in dinoflagellate species diversity at this time reflects a change in depositional environments resulted from a possible regression during a relative stillstand in sea level and subsequent renewed transgression.

Dinocyst Preservation and Lithotypes

As with other kinds of organic matter, dinocysts are preferentially concentrated in fine grained sediments because they are hydrodynamically equivalent to clay and silt grade

clastics (Cross et al., 1966; Davey 1971; Wall et al., 1977). Palynomorphs (including dinoflagellate cysts) are approximately of coarse silt size but of low density. Hence they act as particles of medium silt or finer (Muller 1959) and are deposited with particles of this size (Davey 1971). Coarser grained sediments tend to be dinocyst-poor because their better porosity and permeability assist in the oxidation of dinocysts. The concentration of palynomorphs in sediments varies from near zero in well-sorted sandstones (even if the sandstone contains much black organic matter) to a couple of million per gram of sediment in some sorts of coal and some organic marine sediments (Traverse 1988).

It has been recognized that almost no dinocysts are found in pure carbonate sediments. The real reason for such an absence still remains unclear. One possible explanation is that most dinocysts are destroyed in oxygenated bottom water conditions during depositional and early diagenetic processes. In contrast, this kind of condition is requisite for formation of carbonate sediments. Relatively larger crystallization size of carbonate grains may be a reason, but it has not been tested. Many carbonate grains, whether of inorganic or organic origin (i.e. oolites, aggregates, interclasts, pellets, and skeletal debris), are relatively coarser (sand size). Large grain size provides more opportunity to oxidize dinocysts and other kinds of organic matter after they are deposited. Most limestones contain very few palynomorphs, and dolomites are

invariably barren, probably because of secondary recrystallization and dolomitization effects (Traverse 1988). In impure carbonate sediments, dinocysts can be preserved in various abundances. The relative abundance of dinocyst seems to be a good indicator for the content of carbonate in sediments.

Detailed study of dinoflagellates through the Western Interior basin indicates a correlation between the concentration of dinocysts and lithostratigraphy. Figure 36 shows that the abundance of dinocysts decreases semi-logarithmically with increasing percentages of carbonate fraction. Most points fall in a wide band except two discrepancies: one is the group with the lowest carbonate content having various concentration of dinocysts. Most of such samples are collected from the *Sciponoceras* zone of the Dallas (T1) section. The deposition rates during that time were very high due mostly to delta formation off of rivers flowing out of the Quachita Uplift and/or tectonic subsidence at the section site (Elder 1987b). Although the effect of lower cyst production due to less vigorous upwelling along the eastern margin of the basin can not be ruled out, the dilution factor is probably the primary cause of such unusually low concentrations of dinocysts. Another discrepancy is found in the upper right corner of the diagram with several points showing both relatively high concentration of dinocysts and high carbonate content. The uncommonly high concentration of

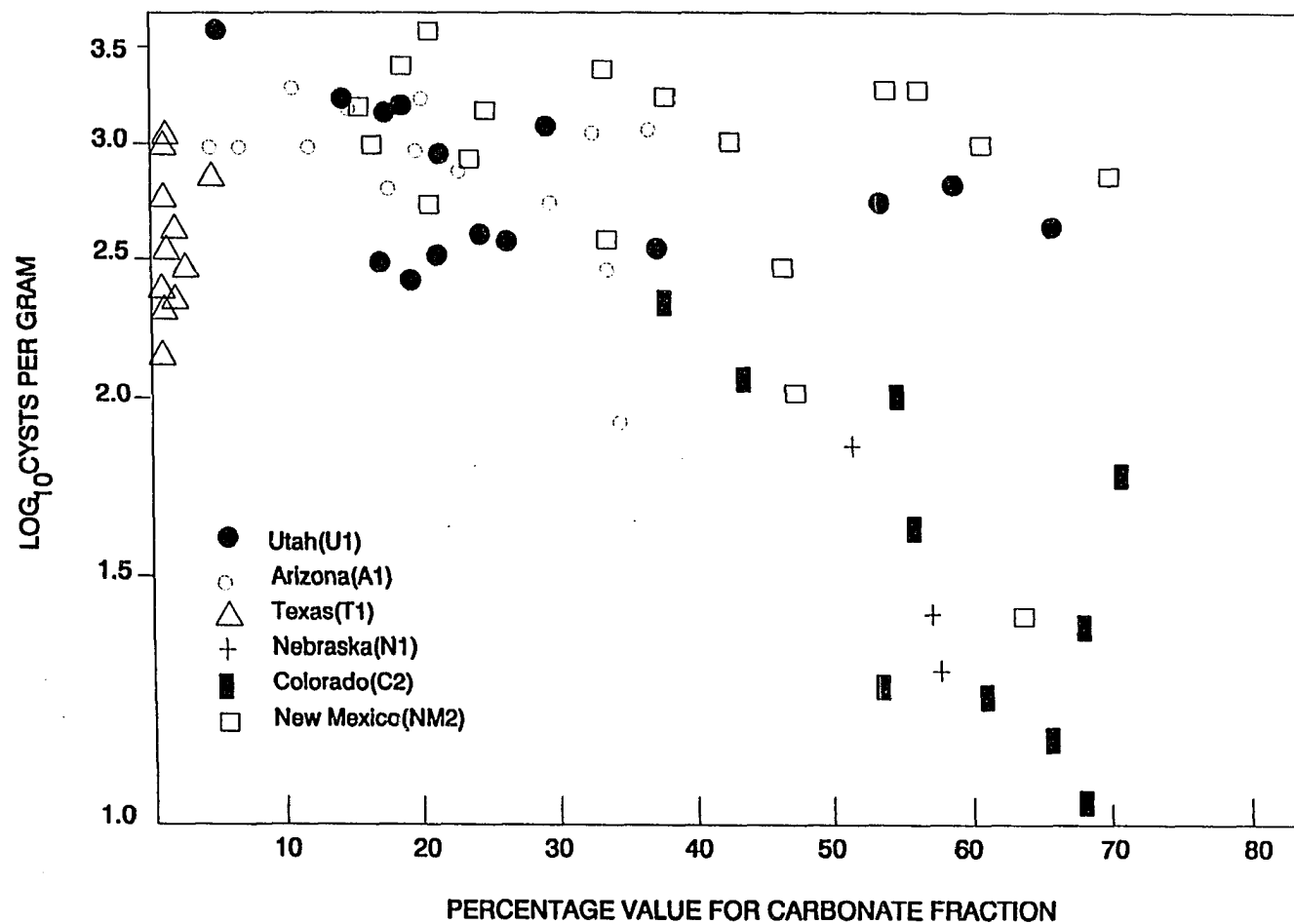


Fig. 36 Relationship between carbonate and dinocyst concentration.

dinocysts from the samples of the *Neocardioceras* zone and the lowest *Watinoceras* zone in the New Mexico (NM2) section (square mark) is probably related to anoxic or anoxic-like environmental conditions for better preservation which is indicated by a peak in organic carbon percentage in such marlstones (Elder 1987b).

Palynological Responses to Sea-Level Changes

With the increasing usefulness of sequence stratigraphic principles and interpretations since the late 1970's, recent studies of organic-rich sediments have integrated stratigraphy, sedimentology, organic geochemistry, organic petrology and palynology in depositional sequence analysis (Habib et al., 1988, 1992; Leckie et al., 1989; Wignall 1991; Pasley and Hazel 1990; Pasley et al., 1991; Gregory and Hart 1992; Moshkovitz and Habib 1993).

Studies of Cretaceous-Tertiary boundary sections in the Atlantic Coast Plain show that the number of dinoflagellate and acritarch species is a very useful criterion for sea level changes (Habib and Miller 1989; Habib et al. 1992; Moshkovitz and Habib 1993). The dramatic increase in species richness above the poorly fossiliferous samples of lowstand intervals implies the beginning of the Transgression Facies Tract during a sea level rise, whereas fewer species indicates either the Lowstand Facies Tract or Highstand Facies Tract. These two

tracts are differentiated by the other abundant organic particles. High concentration of inertinite suggest a coastal environment (lowstand) and abundant amorphous debris an open marine condition (highstand) (Habib et al., 1992; Moshkovitz and Habib 1993). The relationship between facies tract components and the composition of the palynoflora is best represented in study of several subsurface sections of the Gulf Coast (Gregory and Hart 1992). The Lowstand Facies Tract (LFT) is characterized by the dominance of terrestrial-derived particles and high percentages of recycled forms. In the Transgressive Facies Tract (TFT), marine components dominate both in diversity and abundance. Land-plant materials and recycled palynomorphs decrease in number because many of these particles become trapped in estuarine and near-shore environments. The palynoflora in the Highstand Facies Tract (HFT) varies depending upon the proximity to deltas and the stage of the HFT. There is a continuous change from a marine-dominant palynomorph association with relatively few recycled forms in the distal areas (aggradation-dominant) during the early stage to a terrestrial-dominant palynoflora with increasing amounts of recycling materials in the proximal positions (progradation-dominant) during the late stage (Gregory and Hart 1992). Results from the Campanian-Maastrichtian interval in Nigeria show the same relationship between palynofacies and facies tracts (Edet and Nyong 1993). A sedimentary unit characterized by an extensive holomarine

palynofacies is interpreted as transgressive facies tract sediments deposited during a period of relative sea level rise. A unit with fluvio-marine palynofacies constitutes the highstand facies tract sediments deposited during a period of highstand progradation. A non-marine palynofacies unit makes up the lowstand facies tract sediment deposited during sea level fall (Edet and Nyong 1993).

The results of the present investigation(Figs. 37 and 38) are in general agreement with above findings, although a detailed correlation between dinoflagellate data and facies tracts (or parasequence sets and parasequences) can not be drawn from this work due mostly to the unavailability of a well documented study of seismic sequence stratigraphy in the Western Interior seaway. There really is no lowstand facies tract through the study interval (Elder 1993, pers. comm.). The relatively complete cycle is only found in the nearshore area, where the study interval includes the transgressive lag sand which is considered to represent the transgressive surface(Elder 1987b). The transgressive surface marked the base of the transgressive facies tract is recorded below the base of the *Sciponoceras* zone in the offshore sections (Elder 1987b), suggesting an earlier influence of this basinwide marine transgression.

The number of dinoflagellate species increases dramatically at the base of the *Sciponoceras* zone in the nearer-shore sections, which represents the beginning of

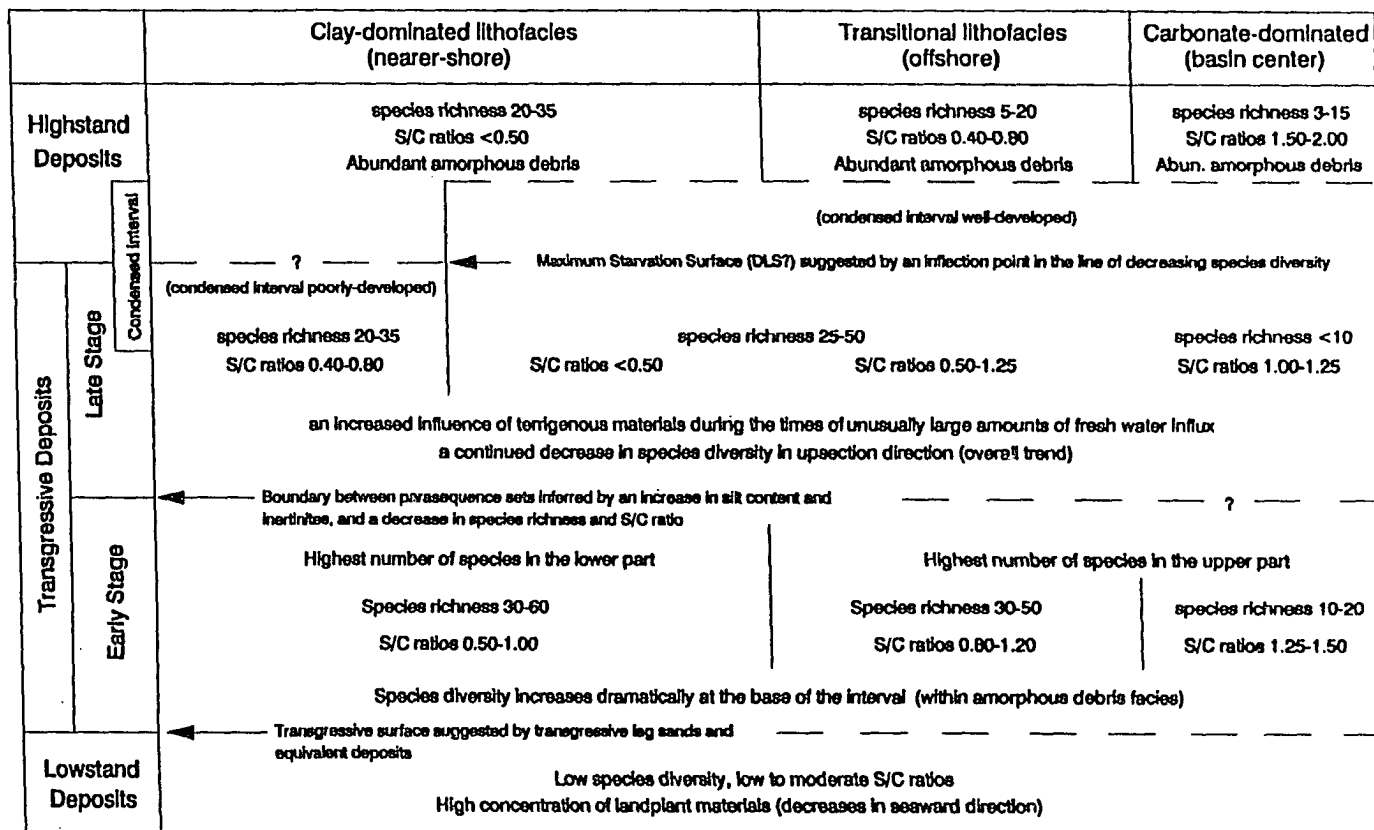


Fig. 37 Composite diagram showing the relationship between the facies tracts and the palynologic results in the three lithofacies regions.

marine transgression. A decrease in species richness associated with an increase in concentration of landplant materials is an indication of the regressive phase during the period of a relative stillstand or a drop of sea level. In the Arizona section, the lower part of the *Sciponoceras* zone has been assigned to the marine transgressive phase and represents a sea-level rise. Species richness increases from fewer than 20 to more than 50. A dramatic decrease in species diversity near the *Sciponoceras-Neocardioceras* zonal boundary is a reflection of regression resulting from a relative stillstand (or a drop) of sea level. The multiple appearances of species immediately above poorly fossiliferous samples at the basal *Neocardioceras* zone indicate the second transgressive pulse.

In the U1 section, the first transgressive phase is not well-documented by the dinoflagellate evidence. Maximum species richness occurs in the lowest sample studied. This is equivalent to the peak interval in the A1 section, suggesting that there may be some less diverse dinocyst assemblages in the samples below during the early transgression. The following regressive phase is indicated by a significant decrease in species richness. The second transgressive pulse at the base of the *Neocardioceras* zone seems to be much less pronounced than that in the A1 section. This may suggest that the effect of this transgressive phase due to a rise of relative sea level is not extensive in the Kaiparowits Basin.

The relative offshore section (NM2) shows a different

reflection in transgression-regression cycle from nearshore area in: (1) the increases in species richness at the base of the *Sciponoceras* zone is much less significant than in the nearshore area. This suggests a lack of environmental change from non-marine to fairly open marine conditions at the beginning of the basinwide transgression; and (2) both the decrease near the *Sciponoceras-Neocardioceras* zonal boundary and the following increase in species diversity are hardly perceived. The smaller effect of marine transgression-regression is probably due to its location farther from the paleoshoreline.

The section in the basin center (C2) shows a similar pattern of dinoflagellate species diversity with the offshore NM2 section. There is an increase in the *Sciponoceras* zone and a following decrease at the *Sciponoceras-Neocardioceras* zonal boundary, although in a smaller scale. The second transgressive pulse is difficult to recognize based on the palynologic data.

Figures 37 and 38 show the conclusion drawn from above that species richness increases dramatically at the beginning of the transgressive phase, then decreases systematically into the highstand interval. The largest number of dinoflagellate species occurs at the base of transgressive intervals (late Cenomanian *Sciponoceras* zone), not in the peak transgression (early Turonian *Watinoceras* zone). The time when the number of species reaches its maximum is different in the nearer-shore

area and in the offshore area (including transitional and carbonate-dominated regions). The earlier time (lower part of the *Sciponoceras* zone) associated with an apparently larger increase in the nearer-shore area suggests that the influence of a rapid expansion of continental shelf at the beginning of a marine transgression is more significant in coastal areas. This coincides with the interpretation of Elder, among many others, that a greater diversification during the early transgressive pulse (*Sciponoceras* zone) in the nearshore area is due to a greater change in depositional environment from essentially non-marine to marine conditions. In contrast, the offshore sections show a characteristic of relatively stable open marine conditions at this time (Elder 1993, pers.comm.). The subsequent decrease in species diversity and increase in inertinite concentration in the nearshore area indicates a regressive interval possibly in response to a stillstand or drop of sea level. The first parasequence set (*Sciponoceras* zone) is well represented by a complete cyclicity in change of species diversity. In the offshore area, the number of species increases gradually during the early stage of the transgressive interval. Then, after a small decrease, species diversity increases again at the beginning of the late transgressive stage. Both increase and decrease are associated with a constant amorphous debris matrix, suggesting a more stable environment during the cycle of transgression and regression. Most sections show a generally decreasing trend in

species richness from the early stage of the transgressive interval through the highstand interval. It is interesting that there is an inflection in the decreasing trend in all the A1, NM2 and C2 sections from which the rate of the decrease in species richness increases. They occur either in the upper part of the *Neocardioceras* zone (A1 and NM2), or in the middle part of the *Watinoceras* zone (C2) which are all considered to be within a condensed, more or less, interval (Elder 1987b, 1993 pers.comm.). There is no such inflection point in the section where a condensed interval is poorly developed (U1 section). The study of Braggs K-T boundary section, Alabama (Habib et al., 1992) showed that the rate of decreasing trend changed within a condensed section. There seems to be a direct correlation between the existence of a turning point in the trend of species richness and the development of a condensed section.

Condensed intervals were defined for thin marine pelagic to hemipelagic sediments formed at extremely low-sedimentation rates during a relative sea-level rise and abrupt transgression of the shoreline (Loutit et al., 1988). Condensed sections are areally most extensive at the time of maximum regional transgression, and are generally considered to be deeper water in origin (Loutit et al., 1988), although some shallow-water type condensed sections have also been recorded (Leckie et al., 1989; Powell 1992). The base and top of a condensed section are dependent on shelf position, sediment

influx, and rate of eustatic sea level change (Baum and Vail 1988). In the nearshore areas, a relatively thin layer of pelagic or hemipelagic sediments is immediately followed by a prograding highstand systems tract which is characterized by a coarsening- or shoaling-upward trend and an upward increase in terrigenous sediment content (Loutit et al., 1988). In the distal parts of the basin, sedimentation rate is still very low at the beginning of prograding accumulation, representing the continuation of the condensed section. The condensed sections generally span the uppermost part of the transgressive facies tract and the lowermost part of the highstand facies tract, with a boundary defined by the downlap surface (Loutit et al., 1988). The seismic downlap surface represents the maximum flooding surface or the starvation surface (Loutit et al., 1988; Baum and Vail 1988). The position of the maximum flooding (starvation) surface within a condensed section varies. It can develop at the base of a condensed section (Donovan et al., 1988). In this case, the condensed section is actually a part of the highstand facies tract (Donovan et al., 1988; Loutit et al., 1988). A condensed section can be seismically identified at the base of major progradational clinoforms of the highstand facies tract. Therefore, the downlap surface was thought to be a good indication of the presence of a condensed section (Loutit et al., 1988). However, the downlap surface is represented by an omission surface in many outcrop sections, and it has been

difficult to observe the downlap of prograding highstand clinoforms in outcrop sections (Loutit et al., 1988).

The similarity of the results from this investigation and Habib's study (1992) leads to the conclusion that the presence of an abrupt change in trend of decreasing species diversity is consistent with the development of a condensed section, which may indicate the position of the maximum flooding (or starvation) surface delimiting the boundary between the transgressive and the highstand facies tracts. The reason for this change is unknown. It may be related to the increased terrigenous dilution as the result of the increased shoreline progradation.

Based on the reading of Elder's papers, it seems to be that the position of a condensed interval within the C-T sequence shifts higher in the offshore direction (Fig. 39). In the nearshore A1 section, the condensed interval extends from LS6 (middle *Neocardioceras* zone) to somewhat above LS7 (Uppermost *Neocardioceras* zone) (Elder 1987b). The condensed interval within the offshore NM2 section is from above LS6 (middle upper *Neocardioceras* zone) to LS8 (lowermost *Watinoceras* zone) (Elder 1987b). In the basin center (C2), the condensed interval appears to start at LS9 (middle *Watinoceras* zone) and extend somewhat above LS11 (uppermost *Watinoceras* zone) (Elder 1993, pers.comm.). Similarly, the position of species diversity inflection, which is used as a criterion of delimiting the boundary between the transgressive and

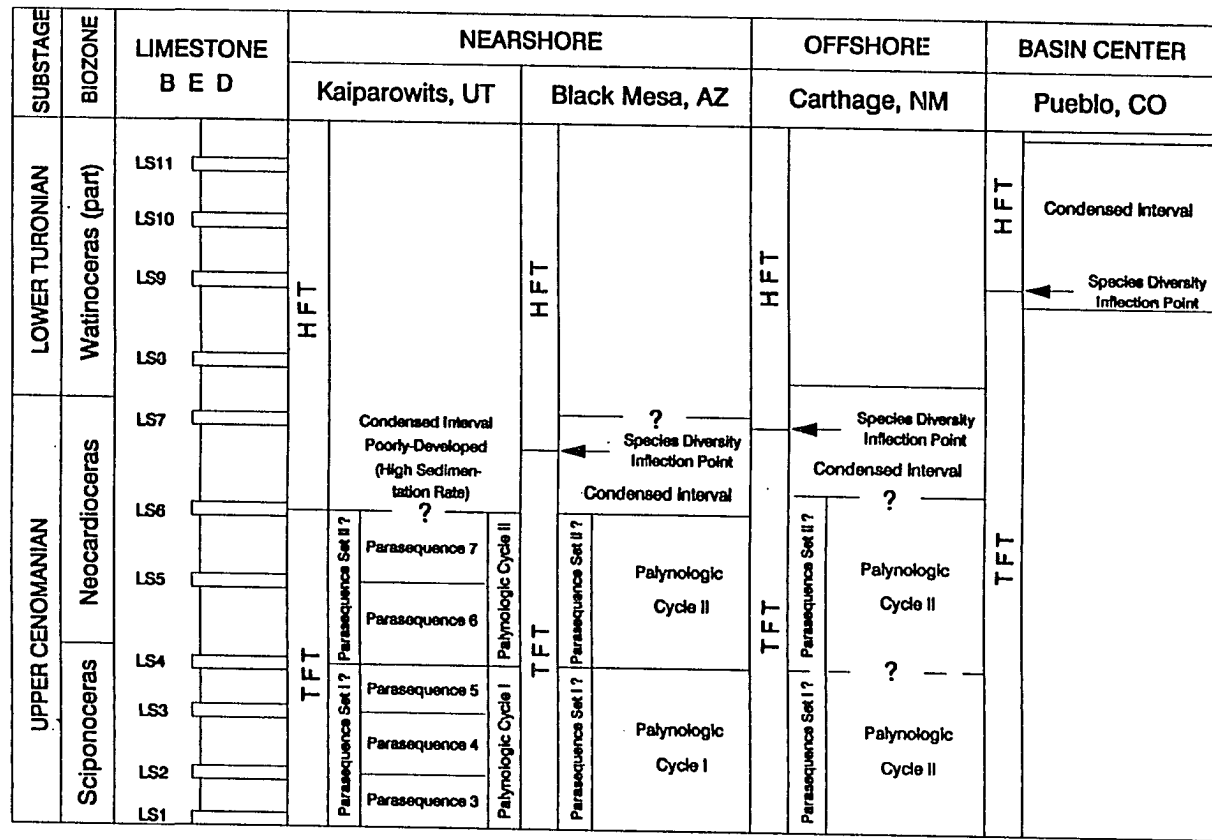


Fig. 39 Composite diagram showing the position of the boundary between the transgressive and highstand facies tracts, which is suggested by the change of dinocyst species diversity. The position

shifts higher in the offshore direction, representing the lagging effect of shoreline progradation in the basin center. Parasequence data from Elder et al. 1994.

highstand facies tracts, is also trending higher away from the shoreline. This may suggest a lagging influence of shoreline progradation in the basin center.

The ratio of *Spiniferites* to *Cyclonephelium* (S/C ratio) is used in the study of the well-developed dinoflagellate assemblages as a supplementary parameter to represent the palynologic response to sea level change. In the nearer-shore area (A1), the increases in S/C ratio and in species diversity, associated with abundant marine-origin amorphous fecal pellets, occur at the beginning of marine transgression. The significant decrease in S/C ratio and the temporary disappearance of most species near the *Sciponoceras-Neocardioceras* zonal boundary indicate a regressive phase. The subsequent changes (increase, and then decrease) of S/C ratio within the latest Cenomanian *Neocardioceras* zone, correspondent with a similar change of species richness, are indicative of the second transgression-regression cycle, although the regressive pulse is of a smaller scale. The magnitude of change in S/C ratio represents, more or less, the scale of transgression and regression. Like species richness, the highest S/C ratio also occurs at the basal transgressive interval. It is followed by a decrease, more or less, through the continuation of marine transgression. This suggests that more open-marine preferential long-process *Spiniferites*-type cysts increase more rapidly than the near-shore habitant *Cyclonephelium*-type cysts during the expansion of shelf

environments. In the relative offshore areas (NM2), the curve of S/C ratio also shows a correspondence with that of species richness. The relatively smaller change in species diversity near the *Sciponoceras-Neocardioceras* zonal boundary makes S/C ratio a unique palynologic criterion that can be used to indicate the transgressive and regressive pulses. However, the significant increase in S/C ratio at the beginning of the transgressive phase is hardly seen in this area which reflects less influence of this sea level change at this offshore position. Relatively high S/C ratios through most of the *Sciponoceras* and the *Neocardioceras* zones suggest that relatively large amounts of *Spiniferites*-type cysts had already existed in the Carthage area during that time.

Facies (systems) tracts (see Friedman et al., 1992) are composed of parasequences or parasequence sets which are relatively conformable succession of genetically related beds or bedsets bounded by marine-flooding surfaces and their correlative surfaces (van Wagoner et al., 1988). Both parasequences and parasequence sets represent shorter-term, higher frequency flooding events which may occur in all systems tracts (Haq et al., 1987). It is difficult to differentiate these two short-term depositional phenomena when a distinctive stacking pattern is not recognized. Based on study of biostratigraphy and cyclostratigraphy, Elder and others (1994) identified seven parasequences (*M.mosbyense*, *S.gracile*, and *N.juddii* zones) in the coarsening-upward

progradational strand plain deposits in southwestern Utah, which may be correlated with the carbonate-dominated (basinal) sequences in the basin center (Colorado). In the present study, the smaller-scale transgressive-regressive palynological cycles within the transgressive interval are represented by dinoflagellate incursions associated with flooding surfaces followed by progradation of terrigenous detritus and fewer dinoflagellate species. They are of larger-scale than those shown by parasequences in the study of Elder and others (1994). The finer palynologic cycles which may represent parasequences are restricted by the broad sampling intervals in the recent study. Each palynologic cycle corresponds two or three parasequences, with cycle I containing three parasequences (parasequence 3 through parasequence 5), and cycle II equivalent to parasequence 6 and parasequence 7 (Fig. 39). These two palynologic cycles may show the stratigraphic units equivalent to parasequence sets if a distinctive stacking pattern can be recognized.

Palynologic cycle I is developed during the early transgressive interval (*Sciponoceras* zone), represented by a dramatic increase in species diversity and S/C ratio above the transgressive lag sands or equivalent deposits, and a subsequent decrease in the number of species concomitant with an increase in silt content and inertinite. These criteria indicate a complete cycle of marine transgression during a sea level rise and regression during a stillstand or drop of sea

level, especially in the nearer-shore areas. Palynologic cycle II represents the deposits in the late transgressive phase (*Neocardioceras* zone). The clearer cyclicity of species richness and S/C ratio is only recorded in the nearshore A1 section. The lower boundary of cycle I has been determined by the presence of transgressive lag sands and equivalent deposits, which indicates the beginning of the transgressive interval (Elder 1987b). The boundary between two cycles (adjacent the *Sciponoceras-Neocardioceras* zonal boundary) is indicated by a significant drop in species diversity, lower S/C ratios, increased inertinite concentration, and an increase in silt content. They all suggest a relative stillstand of sea level. The upper boundary of cycle II (boundary between transgressive and highstand facies tracts) is difficult to determine due to a lack of well-identified downlap surface. However, an inflection in the line of species-number within the condensed interval for all the A1, NM2 and C2 sections is used as an indicator of the boundary between these two facies tracts. In the areas with poorly-developed condensed interval, such as the U1 section, it is impossible even to make such a delimitation since the number of species only decreases gradually in that interval.

The striking correlation between dinoflagellate assemblages and marine transgression-regression cycles provides insight into a framework of sequence stratigraphy. However, any meaningful integrated explanation can not be

successful without a detailed study of seismic stratigraphy.

Paleoceanographic Implications

Numerous methods of investigation have been used to attempt to reconstruct the paleoceanography of the Cenomanian-Turonian Western Interior seaway. Lines of evidence include stable isotopes, sedimentary petrology, macrofossil and microfossil paleoecology, and stratigraphic reconstruction. Much attention has been focused on the origin of the rhythmic bedding in the Bridge Creek Limestone Member of the Greenhorn Limestone and its equivalent strata and the high organic carbon content. At present, two differing hypothesis have been proposed to explain the occurrence of this cyclic bedding in the Cenomanian-Turonian sediments: the dilution model of Pratt (1981) and productivity model of Eicher and Diner (1985, 1989). The evidence for each and their implications on the distribution of organic matter are briefly discussed below.

Pratt (1984), Barron et al. (1985) and Arthur et al., (1985, 1987) among others, have pointed out the shifts in oxygen isotopes in Cenomanian-Turonian cyclic sediments. Although most agree that some diagenetic shift has occurred in the isotopic values, the relative changes in isotopic ratios cannot be explained solely by post-depositional processes. Based on certain assumptions on the isotopic composition of Cretaceous seawater and continental runoff, Arthur et al.

(1985) have postulated a reduction in Greenhorn Sea (Cenomanian-Turonian Western Interior seaway) surface water salinity of as much as 22‰ during peak marlstone deposition relative to chalk deposition during some cycles. The postulated depressed surface water salinity is related to increased terrestrial runoff into the basin during black shale and marlstone deposition (Pratt 1981, 1984; Arthur et al., 1985, 1987). These studies emphasize the importance of freshwater influx in controlling the sedimentary particles available for deposition, degree of bottom oxygenation, benthic population, and storage of organic material. The dilution hypothesis explains the rhythmicity of the Greenhorn cycles by cyclic variations in climates. In general, dry periods are correspondent with relative open circulation within the basin, resulting in the deposition of well-bioturbated, organic-poor limestones. Wet periods are associated with greater runoff, development of salinity stratification of the water column, depauperate benthos, increased storage of organic carbon, and marlstones and black shale deposition. Climatic modelling studies have suggested that rhythmic bedding exhibited in the Greenhorn carbonates can be attributed to climatic fluctuations forced by Milankovitch orbital variations (e.g., Barron et al., 1985). The vertical stagnation of the water column necessary for the storage of large amounts of organic carbon may also have been facilitated by the encroachment of an oceanic oxygen minimum

zone into the Western Interior basin during maximum transgression (Arthur et al., 1987). This model is more appealing because it offers an explanation for the lithology, mineralogy, geochemistry and macropaleontology of the Cenomanian-Turonian limestone-shale cycles.

A simple review on the transportation, deposition of terrigenous organic particles during marine transgression and regression will be helpful to discuss whether the dilution model is supported by palynological evidence.

Sporomorph assemblages in marine sediments are strongly affected by hydrodynamic sorting processes which fractionate the original sporomorph populations according to size, density and morphotype (e.g., Heusser and Balsam 1977; Traverse 1988). The most common onshore-offshore trend is the progressive elimination of the larger, denser cuticular and tracheal tissues and some strongly ornamented pteridophyte spores. This coincides with a gradual increase in the ratio of bisaccate pollen to simple unornamented spores (Hughes and Moody-Stuart 1967; Habib 1979, 1982; Tyson 1985). The relative abundance of terrestrial organic matter is controlled mostly by distance from the shoreline and the direction and strength of marine currents. In coastal areas, especially near river mouths, a substantial contribution of the organic matter buried in the sediments is of terrigenous origin (e.g., Deuser 1972). The terrestrially derived organic matter is generally a very minor component in sediments deposited beyond the shelf-slope break,

even off deltas of large rivers (Jones 1987; Hay 1988). However, large amounts of terrestrially derived organic matter may be carried over the shelf edge and deposited in deep-water marine environments if slumping and turbidity currents are present (Cornford 1979; Tissot et al., 1980). Pollen grains also can be carried by wind or insects, but the amount is very small compared to the total production of pollen/spores (Traverse 1988). Many of such particles normally settle within a kilometer of the source plant. Once spores/pollen grains reach the ground, further movement is mostly by water (Traverse 1988). The study of the Orinoco Delta shows that a considerable amount of airborne pollen is present in the sediments, but it is numerically overwhelmed by water-borne pollen (Muller 1959).

Large amounts of freshwater runoff from a humid vegetative highland should lead to the deposition of unusually large amounts of terrestrial organic matter in coastal and shelf areas, if no strong circulation exists in the basin. The case of the Black Sea has been considered as a typical example of the anoxic bottom conditions resulted from a brickish water cap due to high terrigenous influx (Hay and Honjo 1989). Recent data coming out of organic particle sedimentation studies show that the laminated carbonaceous sediments widespread in the Black Sea are comprised largely of landplant detritus evidently contributed by streams discharging into the basin (Hay and Honjo 1989). High terrigenous matter

accumulation rates suggests high precipitation in the drainage area of the rivers surrounding the Black Sea (Hay 1988). Some of the laminated sediment sequences can extend over nearly 1000 km (Honjo et al., 1988). The removal of terrigenous matter occurred dominantly during the intense plankton bursts in spring and summer/fall, with fecal pellets and 'marine snow' aggregates as the main transporting agents (Hay and Honjo 1989). Nutrients are mainly supplied to the surface water by upwelling of the nutrient-rich deeper water and by riverine nutrient supply. Therefore, higher primary production is also expected closer to shore (Glenn and Arthur 1985).

Recent studies of sequence stratigraphy have demonstrated that depositional facies active during times of sea level highstand (Highstand facies tract) and lowstand (lowstand and shelf margin facies tracts) are influenced heavily by progradation. This results in the introduction of considerable amounts of terrigenous sediments and associated terrestrially derived organic matter to the shelf environment. Whereas depositional facies active on the shelf during transgression of the shoreline (transgressive facies tract) receive relatively little terrigenous sediment because most of this material is trapped in estuarine and other coastal environments (Swift 1976; Dominques et al., 1987; Jervey 1988; Loutit et al., 1988; Posamentier and Vail 1988; Roberts and Coleman 1988; Leithold and Bourgeois 1989). Therefore, shelf sediments in the transgressive facies tract should contain a

higher percentage of autochthonous (marine) organic matter. Studies of modern estuaries have proven them to be efficient sediment traps (MaCave 1973; Nichols and Biggs 1985). Of special interest to this discussion is the work of LeBlanc et al. (1989) who found that the amount of terrestrial organic matter decreases steadily and rapidly seaward in the estuarine environments of Negro Harbor, Nova Scotia, and that during transgressions, terrestrial organic matter is trapped in estuarine environments and is probably deposited on the shelf only after spending considerable time exposed to bacterial and fungal degradation. Leckie et al. (1990) have shown that the reduction in terrestrial palynoflora may be due to the inundation of a large area by transgression and the wholesale destruction of the regional vegetation. Alternatively, a rapid inundation of the continental land mass can also drastically reduce the terrigenous debris, including the palynoflora, entering the sedimentary basin. Most phytoclasts in the transgressive facies tract have undergone extensive degradation presumably brought about by lower sedimentation rates. Conversely, organic matter deposited in the highstand facies tract is dominantly terrestrial in origin and exhibits little degradation because of high terrigenous sediment influx and high sedimentation rates associated with progradation.

Many factors, other than type of organic input, control the deposition in marine environments. The residence time of the organic matter in an oxic environment is important,

because the plant tissues undergo much longer exposure to an oxidizing environment than other terrestrially derived organic particles. The composition of organic particles will change radically despite a constant type of organic matter input if the oxygen content of the depositional environment changes (Jones 1987).

Previous palynological studies in some terrigenous interior and marginal basins in the belt along the western shoreline of the seaway have suggested warm temperate to subtropical vegetation on the uplands to the west of the seaway during the Cenomanian-Turonian times (e.g., Upshaw 1964; Pannella 1966; Agasie 1969; Singh 1964, 1971, 1983; Nichols and Jacobson 1982a, 1982b). Climatic modelling studies have predicted the Cretaceous warmth and geography-promoted high global precipitation rates, particularly in the northern hemisphere mid-latitude continents (Barron and Washington 1982; Parrish and Barron 1986). The widely distributed coals and kaolinities in North America are very good indicatives of high rainfall during the mid-Cretaceous time (Beeson 1984; Turtelot 1983; Parrish and Barron 1986).

Assuming that the dilution hypothesis is right, one would expect abundant landplant material, especially some pollen grains and spores, in nearshore sediments as a result of high quantity of fresh water influx from the western humid, vegetative highland. However, it is difficult to fit all evidence of palynomorphs and other organic matter into this

model. The palynologic studies for the two nearer-shore sections (A1 and U1) exhibit well-preserved palynomorph assemblages with diverse dinocysts, very low ratios of sporomorph species to dinoflagellate species, and a low concentration of terrestrially-derived organic matter. The overall trend does not change much even in the intervals of mixed facies of inertinite and amorphous debris. This is much different from the case of Black Sea. Both A1 and U1 sections are believed to be located 100-150 km from the tectonically active western margin of the seaway (Elder 1987b). This is much shorter distance than some 1,000 km-extensive organic-rich sediments in the Black Sea. In general, even 100 km is too far for landplant materials to be carried only by riverine transportation. Most of such particles would be deposited in prodelta and nearby areas. But considering similar climatic and geographic conditions, such as warmer weather, higher precipitation, larger amounts of fresh water runoff, and shallower and gentle-slope epicontinental seas, one still would expect to find more landplant particles no matter how they are preserved (complete or only fragmented). There is no doubt that landplant materials, particularly bisaccate pollen grains and some spores, can be carried at least same distance as terrigenous clay minerals if they are not totally destroyed during transportation processes due to oxidation. In light of the palynological evidence, the effect of continental runoff resulting in salinity stratification of the water column

should be viewed as somewhat overestimated.

Foraminiferal evidence that there are no apparent differences between the planktonic foraminiferal assemblages from the limestones and marlstones has led Eicher and Diner (1985) to propose a different model for the origin of the rhythmic bedding in the C-T strata. They proposed that carbonate sediments were deposited during dry periods when warm saline bottom water (WSBW) formation is increased. Vertical circulation would be enhanced and resulted in upwelling and increased productivity. Black shales and marlstones were deposited during wet periods when WSBW formation in subtropical epicontinental sea is much reduced, and the circulation in the western interior basin would be sluggish. This hypothesis depends on oceanic circulation and primary productivity resulting from upwelling. Study on the distribution of dinoflagellate cysts and other organic matter might provide a good test for oceanic circulation, but generally fail to test primary productivity. In most circumstances the level of primary productivity is strongly influenced by the efficiency of nutrient recycling within the ecosystem. Shelf productivity is often based on efficient upwelling recycling, which is thought to be a greater nutrient source than, for example, river water (Rowe et al., 1975). Dinocyst density may be partly related to productivity (Davey 1971; Davey and Rogers 1975), but there is no consistent general correlation between dinocyst abundance and the

productivity of the overlying waters. Moreover, it is very possible that even in areas of relatively low productivity (e.g., Cretaceous Western Interior seaway), well-represented and abundant dinocyst will accumulate if diagenetic and preservational factors are fulfilled. The redeposition of dinocysts, acritarchs and other organic matter may in part reflect the extent of oceanic circulation and shelf upwelling. The general lack of reworked dinocysts recognized by the discordant stratigraphic appearance, and a minor recycling of amorphous debris distinguished by dark color, poor preservation and its association with inertinite (Habib et al., 1994) in most samples studied suggest that circulation in the Western Interior seaway may have been sluggish during the wet periods when black shales and marlstones were deposited.

Both of these models may be applicable to some extent. A couple of other factors should also be taken into account for explaining the stratigraphic distribution of terrestrially derived organic materials in the Cenomanian-Turonian Western Interior seaway. The paucity of sporomorphs in most samples including the nearshore sequences, and poor preservation of tracheal and other cuticle tissues indicate that large amounts of landplant materials may have been oxidized and degraded prior to or during transportation in subaerial and well-oxygenated submarine conditions before final settling to the bottom of the basin. The maximum depth in the basin center at the peak transgression was estimated to be 500-600 m (Kauffman

1977). However, much of the seaway may have been as shallow as 100 m during sea-level lowstands (Arthur et al., 1987). The relatively shallow depth might lead to well-oxygenated bottom conditions because of effective wind and wave mixing, which provides the potential for landplant oxidation and degradation. Different kinds of landplant particles have various degrees of resistance to the oxidation. Cuticular plant debris is much less refractory, and tends to be concentrated in pro-delta sediments opposite the mouths of large distributaries (Muller 1959; Cross et al., 1966). Pollen grains and spores are relatively refractory and can survive long enough to be potentially widely distributed. The absolute abundance of terrestrial organic matter is considered as mainly a function of the climatic controls on the parent vegetation and the proximity to fluvial inputs (Tyson 1987). Since vegetation in the highland source area was extensive during episodes of wet climate, and large river discharge was produced, the rarity of terrestrial plant particles could be partly attributed to the relatively distal positions to fluvial inputs. It is apparent that terrestrial particles brought by rivers from highlands would be mostly deposited in delta and nearby areas if bottom circulation is sluggish in the basin. The relatively short distance away from shoreline does not necessarily mean the relatively high content of terrestrial organic matter except in the immediate proximity of fluvial sources. The stratigraphic and sedimentologic

results from the southwest Utah area suggests that nearshore sediments were derived from the Mogollon Highlands (southwest Arizona) to north (southeast Utah) during the studied interval (Gustason 1989). However, similar concentrations of land-plant materials in both the Arizona (A1) and Utah (U1) sections and even less degraded landplant material in the U1 than in the A1 section indicate that the movement of sediment via longshore drift is not as strong as Gustason's data suggested.

Neither the dilution nor productivity models for the Cenomanian-Turonian rhythmically-bedded sediments can solely explain the distribution of dinoflagellate cysts, sporomorphs and other organic matter. The influence of brackish surface water cap resulting from high river discharge might be not as extensive as suggested in the dilution model. The bottom conditions during the deposition of the C-T sediments may not be exclusively anoxic, at least in coastal areas (Zelt 1985; Elder 1987b, 1989, 1991). This provides the potential of oxidation of most less refractory land plant particles. The scarcity of refractory pollen grains and spores in most sequences may reflect the limited effect of fresh water influx. The anoxic bottom environments necessary for the preservation of organic matter during the deposition of black shales and marlstones may be better explained by the joint effect of fresh water influx in the nearshore areas and sluggish circulation in the basin center.

CONCLUSIONS

Rhythmically-bedded organic-rich sediments of the Cenomanian-Turonian Greenhorn Limestone and stratigraphic equivalents contain abundant, well-preserved dinoflagellate assemblages which provide valuable insight into the interpretation of depositional environments. The main conclusions from this study are as follows:

(1) dinoflagellate species diversity, along with the distribution of other organic matter, is closely related to marine transgression and regression resulting from change in relative sea level. The dramatic increase in the number of species within amorphous debris facies is considered as the best palynological evidence for marine transgression. The decrease in species diversity and low ratio of dinocyst to sporomorph species defined by an admixture of inertinite and amorphous debris facies represents a regressive pulse. The first transgressive pulse within the study interval occurs basinwide at the base of the *Sciponoceras* zone, with a significant increase in species diversity associated with abundant amorphous debris. This is followed by a marked drop in the number of species associated with an increase in inertinite content adjacent the *Sciponoceras-Neocardioceras* zonal boundary, which is interpreted to represent a regressive

interval resulting from a stillstand or possible drop of sea level. The second marine transgression is characterized by the multiple occurrence datums immediately above poorer assemblages in the basal *Neocardioceras* zone. This increase in dinocyst species is not as pronounced as the first, and is only locally developed. A decrease in species diversity adjacent the C-T boundary is only recorded in the nearshore A1 section. This less apparent and smaller change in species richness in the *Neocardioceras* zone reflects a diminished effect of transgression-regression cycle resulting from a smaller fluctuation of relative sea level or possibly a further position offshore as transgression continued. The cyclic change in species diversity in response to marine transgression and regression is only recorded in the western nearer-shore sequences.

(2) The paleoecological significance of *Spiniferites* and *Cyclonephelium* based on previously studies, and their inverse relationship make them a very useful in interpretation of depositional environments. The ratio of *Spiniferites* to *Cyclonephelium* (S/C ratio) is calculated in most sections as a supplementary parameter to species diversity and depositional organic facies. As does species richness, S/C ratio increases in the transgressive interval (more apparent at the beginning), and decreases in the regressive pulse. It is particularly useful in smaller-scale transgression-regression cycles, where there is little change in species

diversity. The invariably higher S/C values in the equivalent horizons in the NM2 section than in the A1 section indicate the former is deposited in a more open marine shelf condition.

(3) Among all major types of organic matter (both in marine and terrestrial origin) deposited in the C-T sediments, amorphous debris, mostly formed as feces produced by zooplankton consuming organic-walled microplankton (Habib et al., 1992), predominates in all sequences. It constitutes approximately 80 percent or even higher of all organic particles. Landplant remains, including pollen grains and spores and various tissues, are a minor component in most samples. They show a slight increase in the *Neocardioceras* zone in response to the increased freshwater influx. Inertinite is important in the nearer-shore areas in relation to the change in depositional environments mostly resulted from marine regression. Dinoflagellate cysts and acritarchs are also a minor component. Their diversities vary with the changes of concentration of amorphous debris and inertinite, especially in the nearer-shore areas at the beginning of the environmental changes.

(4) This palynological study can be interpreted in a sequence stratigraphic context involving the reconstruction of sea level curves. The number of species is lower at the very beginning of the transgressive interval, representing a nearshore environment. It increases pronouncedly at the beginning of the transgression (basal *Sciponoceras* zone) as the

result of a quickly expanded shelf environment during a rise of sea level. The highest number of species occurs in the lower part of the transgressive interval, directly above the poorly fossiliferous deposits in the nearshore areas, but a little bit later in the relatively offshore environments. This suggests a more marked influence of shelf expansion during a rapid transgression in the coastal and nearshore areas. Species richness remains relatively high particularly in some relatively offshore sections but decreases systematically into the highstand interval, indicating an increased influence of shoreline progradation. An inflection point in the line of decrease in species diversity is correlated with the development of a condensed interval, and interpreted to possibly delimit the boundary between the transgressive and highstand facies tracts. The position of the inflection point shows the same change with the position of the condensed interval that shifts higher in the offshore direction within the overall C-T transgressive sequence. The shorter cycles of transgression and regression, represented by palynologic cycles, may indicate flooding intervals corresponding to a higher sequence unit than parasequence, which is better called parasequence sets if there is a distinctive stacking pattern. The bottom boundary of cycle I (*Sciponoceras* zone) is marked by the transgressive surface. Cycle I is separated from cycle II (*Neocardioceras* zone) by a decrease in species richness and S/C ratio, and an increase in inertinite and landplant

materials near the *Sciponoceras-Neocardioceras* zonal boundary. The upper boundary of cycle II is drawn by the inflection point in the line of the number of species in the condensed interval.

(5) Unlike most macrofaunas, dinoflagellates do not show any level of extinction in the Cenomanian-Turonian boundary interval. The temporary disappearances near the *Sciponoceras-Neocardioceras* zonal boundary and prior to or at the C-T boundary, more apparent in the nearer-shore areas, are thought to be the reflection of unstable environments mostly resulted from sea level stillstands.

(6) The preservation of dinoflagellate cysts is highly variable in the sections investigated and closely corresponds to lithology. The best preservation was observed in the calcareous and marly shales of the clay-dominated lithofacies region. In the carbonate-dominated lithofacies area, many samples proved to be devoid of dinocysts or are poorly preserved. There is a close relationship between the dinoflagellate abundance and the carbonate content in strata. Detailed analysis from most samples shows that the concentration of dinocysts in the C-T sediments decreases semi-logarithmically with increasing percentages of carbonate fraction.

(7) The concentration of terrestrially derived organic matter in most samples is much lower than Recent Black Sea sediments. Both sediments have been interpreted to be

deposited under a similar shelf condition with the similar influence of weather, precipitation, hinterland vegetation and freshwater runoff. This may suggest that in the Cretaceous Western Interior basin, either the terrigenous input from the western hinterland is not as high as other data indicated, or a considerable amount of terrestrial organic matter was oxidized and degraded, even destroyed in subaerial and well-oxygenated submarine environments prior to or during transportation processes. The gentle slope and relatively shallow depth over much of the Western Interior seaway appear to have led to the development of dysaerobic, not exclusively anoxic, bottom conditions, which in turn provided the potential for landplant oxidation and degradation.

APPENDIX A
SYSTEMATIC DESCRIPTIONS

Achomosphaera ?neptuni (Eisenack 1958) Davey & Williams
1966

Previously reported occurrences. early Cretaceous.

Achomosphaera ramulifera (Deflandre 1937) Evitt 1963

Previously reported occurrences. late Cretaceous.

Achomosphaera sagena Davey & Williams 1966

Previously reported occurrences. Cenomanian.

Apteodinium granulatum Eisenack 1958

Previously reported occurrences. Aptian.

Ascodinium acrophorum Cookson & Eisenack 1960

Previously reported occurrences. Late Albian-Cenomanian.

Ascodinium parvum (Cookson & Eisenack 1958) Cookson &
Eisenack 1960

Previously reported occurrences. late Albian-Cenomanian.

Batiacasphaera euteiches (Davey 1969) Davey 1979

1969 *Chytroeisphaeridia euteiches* Davey, p.141, pl.3, fig.8-9.

1979 *Batiacasphaera euteiches* (Davey) Davey, p.217.

Previously reported occurrences. Cenomanian.

Batiacasphaera reticulata (Davey 1969) Davey 1979

1969 *Chytroeisphaeridia reticulata* Davey, p.14, pl.4, fig.3-4,
6.

1979 *Batiacasphaera reticulata*(Davey) Davey, p.217.

Previously reported occurrences. Campanian-Maastrichtian.

Bourkidinium psilatum Singh 1983

1983 *Bourkidinium psilatum* Singh, p.125, pl.42, figs.8-9,
pl.43, fig.1.

Previously reported occurrences. Cenomanian.

Chytroeisphaeridia baetica Riegal 1974

Previously reported occurrences. ?Senonian.

Chytroeisphaeridia minor(Cookson & Hughes 1964)
Morgan 1980

1964 *Canningia minor* Cookson & Hughes, p.43, pl.8, fig.1-3, 5.

1980 *Chytroeisphaeridia minor*(Cookson & Hughes) Morgan, p.19.

Previously reported occurrences. Albian-Cenomanian.

Cleistosphaeridium aciculare Davey 1969

Previously reported occurrences. Cenomanian.

Cleistosphaeridium clavulum(Davey 1969) Below 1982

1969 *Cleistosphaeridium polypes* var. *clavulum* Davey, p.154,
pl.6, fig.9-10.

1982 *Cleistosphaeridium clavulum*(Davey) Below, p.15.

Previously reported occurrences. Cenomanian.

Cleistosphaeridium huguoniotii(Valensi 1955) Davey 1969

1955 *Hystrichosphaeridium huguoniotii* Valensi, p.38, text-fig,
2a.

1969 *Cleistosphaeridium huguoniotii*(Valensi) Davey, p.135.

Previously reported occurrences. late Cretaceous.

***Cleistosphaeridium ?multispinosum*(Singh 1964) Brideaux
1971**

1964 *Baltisphaeridium ?multispinosum* Singh, p.141-142, pl.20,
fig.1-2.

1971 *Cleistosphaeridium ?multispinosum*(Singh) Brideaux, p.93.

Previously reported occurrences. Albian-Cenomanian.

***Cleistosphaeridium parvum* Davey 1969**

Previously reported occurrences. Cenomanian.

***Coronifera minus*(Yu & Zhang 1980) Mao & Norris 1988**

1980 *Diphyes minus* Yu & Zhang, p.111-112, pl.4, fig.4-5.

1988 *Coronifera minus*(Yu & Zhang) Mao & Norris, p.36

Previously reported occurrences. Cenomanian-Santonian.

***Coronifera oceanica* Cookson & Eisenack 1958**

1958 *Coronifera oceanica* Cookson & Eisenack, p.45, pl.12,
figs.5, 6.

Previously reported occurrences. Berriasian-Paleocene, Canada (Davey 1969, Harland 1973, Williams 1975, Williams & Brideaux 1975, Singh 1983); Cenomanian-Turonian, Arizona(Jones 1976); Campanian-Maastrichtian, New Jersey(May 1980); Maastrichtian-Paleocene, Maryland(Benson 1976); Hauterivian-Aptian, North Atlantic Ocean (Habib 1972); Barremian-Coniacian, England(Cookson & Hughes 1964, Clarke & Verdier 1967, Davey 1969, 1974); Hauterivian-Cenomanian, France(Davey 1969, Millioud 1969, Davey & Verdier 1971); Aptian, Germany (Eisenack 1958); Aptian-Campanian, Australia(Cookson & Eisenack 1958, 1968, 1969, Norvick 1976, Morgan 1980); Albian, India(Jain 1977).

***Coronifera striolata*(Deflandre 1937) Stover & Evitt
1978**

1937 *Hystriospheridium striolatum* Deflandre, p.72-73,
pl.12, fig.1-2.

1978 *Coronifera striolata*(Deflandre) Stover & Evitt, p.148.

Previously reported occurrences. Late Cretaceous.

Cyclonephelium compactum Deflandre & Cookson 1955

Previously reported occurrences. Albian-santonian.

Cyclonephelium distinctum Deflandre & Cookson 1955

Previously reported occurrences. Senonian.

Cyclonephelium membraniphorum Cookson & Eisenack 1962

Previously reported occurrences. Albian-Cenomanian.

Cyclonephelium paucispinum Davey 1969

Previously reported occurrences. Cenomanian.

Cyclonephelium vannophorum Davey 1969

1969 *Cyclonephelium vannophorum* Davey, p.168, pl.9, fig.3,
pl.11, figs.11-12.

Previously reported occurrence. Cenomanian, England(Davey 1969); Vraconian, France(Davey & Verdier 1973); Cenomanian-Turonian, Australia(Norvick 1976); Hauterivian-Coniacian, Canada (Williams 1975, Singh 1983); Cenomanian, North Atlantic (Habib 1972).

Diphyes cretaceum Yu & Zhang 1980

Previously reported occurrences. Cenomanian-early Turonian.

Epelidosphaeridia spinosa(Cookson & Hughes 1964)
Davey 1969

1964 *Palaeoperidinium spinosum* Cookson & Hughes, p.49, pl.8,
fig.6-8.

1969 *Epelidosphaeridia spinosa*(Cookson & Hughes) Davey, p.143.

Previously reported occurrences. early Cenomanian.

Exochosphaeridium bifidum(Clarke & Verdier 1967)
Clarke et al. 1968

1967 *Baltisphaeridium bifidum* Clarke & Verdier, p.72-73,
pl.17, fig.5-6, text-fig.30.

1968 *Exochosphaeridium bifidum*(Clarke & Verdier) Clarke et al., p.182.

Previously reported occurrences. Cenomanian-Campanian.

***Exochosphaeridium phragmites* Davey et al. 1966**

1966 *Exochosphaeridium phragmites* Davey et al., p.165, pl.2, figs.8-10.

Previously reported occurrences. Valanginian to Maastrichtian. Cenomanian-Campanian, Canada(Harland 1973, Singh 1983); Barremian-Cenomanian, England(Davey et al. 1966, Davey 1969, 1974); Aptian-Cenomanian, France(Davey 1969, Davey & Verdier, 1971, 1974); Senonian, Italy(Corradini 1973); Cenomanian-Maastrichtian, Europe (Foucher 1979), Neocomian-Senonian, Australia(Cookson & Eisenack 1970, Norvick 1976, Morgan 1980); Albian, India(Jain 1977).

***Florentinia cooksoniae*(Singh 1971) Duxbury 1980**

1971 *Hystriosphraeridium cooksoni* Singh, p.329, pl.51, figs.7-8, pl.52, figs.1-4.

1977 *Hystriosphraeridium cooksoniae* Lentin & Williams, p.81.

1980 *Florentinia cooksoniae*(Singh 1971) Duxbury, p.120, pl.7, figs. 6, 9, text-fig.7.

Previously reported occurrences. Barremian to Cenomanian. Barremian-Cenomanian, Canada(Singh 1971, 1983, Brideaux 1971, 1977, Bujak & Williams 1978); Barremian, England(Duxbury 1980).

***Florentinia deanei*(Davey & Williams 1966) Davey & Williams 1973**

1966 *Hystriosphraeridium deanei* Davey & Williams, p.58, pl.6, figs.4, 8.

1973 *Florentinia deanei*(Davey & Williams 1966) Davey & Williams, p.187, pl.1, fig.9.

Previously reported occurrences. Vraconian-Turonian, France (Davey 1969, Davey & Verdier 1973); middle Cenomanian-Turonian, England (Davey & Williams 1966, Clarke & Verdier 1967, Davey 1969).

***Florentinia mantellii*(Davey & Williams 1966) Davey & Verdier 1973**

1966 *Hystriosphæridium mantellii* Davey & Williams, p.66, pl.6, fig.6.

1973 *Florentinia mantellii*(Davey & Williams 1966) Davey & Verdier, p.187-191, pl.1, figs.1, 4, 7, pl.4, figs.1, 3.

Previously reported occurrences. Cenomanian-Turonian, England (Davey & Williams 1966, Davey 1969); Vraconian-Turonian, France (Davey & Williams 1966, Davey 1969, Davey & Verdier 1973).

***Florentinia radiculata*(Davey & Williams 1966) Davey & Verdier 1973**

1966 *Hystriosphæridium radiculatum* Davey & Williams, p.65, pl.7, fig.9, pl.9, fig.6.

1973 *Florentinia radiculata*(Davey & Williams 1966) Davey & Verdier, p.191-192, pl.4, fig.7.

Previously reported occurrence. Albion-Turonian, France, England (Davey & Williams 1966, Davey 1969, Davey & Verdier 1971, 1973).

***Fromea amphora* Cookson & Eisenack 1858**

Previously reported occurrences. Albion-Cenomanian.

***Gonyaulacysta cassidata* (Eisenack & Cookson 1960) Sarjeant 1966**

1960 *Gonyaulax helicoidea* subsp. *cassidata* Eisenack & Cookson, p.3, pl.1, fig.5-6.

1966 *Gonyaulacysta cassidata*(Eisenack & Cookson) Sarjeant, p.125-126.

Previously reported occurrences. Aptian-Cenomanian.

***Gonyaulacysta helicoidea*(Eisenack & Cookson 1960) Sarjeant 1966**

1960 *Gonyaulax helicoidea* Eisenack & Cookson, p.2, pl.1, fig.4, 9.

1966 *Gonyaulacysta helicoidea*(Eisenack & Cookson) Sarjeant, p.116-117.

Previously reported occurrences. Neocomian-Aptian.

Heterosphaeridium conjunctum Cookson & Eisenack 1968

Previously reported occurrences. Santonian-early Campanian.

Hystriochodinium pulchrum Deflandre 1935

Previously reported occurrences. Senonian.

Hystriochosphaeridium bowerbankii Davey & Williams 1966

Previously reported occurrences. Cenomanian.

Hystriochosphaeridium difficile Manum & Cookson 1964

Previously reported occurrences. Cenomanian.

Hystriochosphaeridium recurvatum (White 1842) Davey &
Williams 1966

1842 *Xanthidium recurvatum* White, p.39, pl.4, fig.12.

1966 *Hystriochosphaeridium recurvatum* (White 1842) Davey &
Williams, p.67-68.

Previously reported occurrences. Senonian.

Hystriochosphaeridium stellatum Maier 1959

Previously reported occurrences. Oligocene.

Hystriochosphaeridium tubiferum (Ehrenberg 1838)
Deflandre 1937

Previously reported occurrences. late Cretaceous.

Isabelidinium magnum (Davey 1970) Stover & Evitt 1978

1970 *Chatangiella magnum* Davey, p.140, pl.11, fig.3-4.

1978 *Isabelidium magnum* (Davey) Stover & Evitt, p.109.

Previously reported occurrences. Cenomanian.

***Kiokansium polypes*(Cookson & Eisenack 1962) Below 1982**

- 1962 *Hystrichosphaeridium recurvatum* subsp. *polypes* Cookson & Eisenack, p.491, pl.4, fig.11-13.
 1969 *Cleistosphaeridium polypes*(Cookson & Eisenack) Davey 1969, p.154.
 1982 *Kiokansium polypes*(Cookson & Eisenack) Below, p.16.

Previously reported occurrences. Albian-Cenomanian.

***Kiokansium williamsii* Singh 1983**

- 1983 *Kiokansium williamsii* Singh, p.150, pl.54, figs.3-6.

Previously reported occurrences. Cenomanian.

***Litosphaeridium siphoniphorum*(Cookson & Eisenack 1958)
Davey & Williams 1966**

- 1958 *Hystrichosphaeridium siphoniphorum* Cookson & Eisenack, p.44, pl.11, figs.6-10.
 1966 *Litosphaeridium siphoniphorum*(Cookson & Eisenack 1958) Davey & Williams in Davey et al., 1966, p.80-82, pl.7, figs, 7, 8, text-fig.16.

Previously reported occurrences. Albian-Cenomanian, Australia (Cookson & Eisenack 1958, 1968); upper Albian-lower Cenomanian, England(Cookson & Hughes 1964, Davey & Williams 1966, Clarke & Verdier 1967, Davey 1969); Albian, Rumania (Baltes 1963, 1967); Albian-Turonian, Canada(Davey 1969); Cenomanian, France(Davey 1969, Davey & Verdier 1973); Albian-Cenomanian, Germany; Vraconian, Morrocco(?). Cenomanian, U.S.A.(Davey 1969); Cenomanian-Turonian, Kansas(Lucas-Clark 1984).

***Odontochitina costata* Alberti 1961**

- 1961 *Odontochitina costata* Alberti, p.31, pl.6, figs.10-13.

Previously reported occurrences. ?Albian-Cenomanian, Australia (Cookson & Eisenack 1962); Cenomanian-Campanian, England (Cookson & Hughes, 1964, Clarke & Verdier 1967, Davey 1970); Vraconian-Cenomanian, France(Davey & Verdier 1973); Cenomanian-Turonian, Germany(Alberti 1961); Upper Albian-lower Cenomanian, Canada (Davey 1970); Upper Cretaceous, Canada(Manum & Cookson 1964).

Odontochitina operculata(O.Wetzel 1933) Deflandre &
Cookson 1955

Previously reported occurrences. Senonian.

Oligosphaeridium complex(White 1842) Davey & Williams
1966

Previously reported occurrences. Senonian.

Ovoidinium verrucosum(Cookson & Hughes 1964) Davey
1970

Previously reported occurrences. Cenomanian.

Palaeohystrichophora infusorioides Deflandre 1935

Previously reported occurrences. Vraconian(latest Albian) to Maastrichtian. ?Albian-Santonian, Germany (Alberti 1961); Cenomanian-lower Turonian, Australia(Cookson & Eisenack 1958); Cenomanian-Turonian, Rumania(Baltes 1966); Cenomanian-Santonian, England(Cookson & Hughes 1964, Clarke & Verdier 1967, Davey 1970); Vraconian-Senonian, France(Deflanfre 1935, 1936, 1940, Davey 1970, Davey & Verdier 1973); Albian-Senonian, Germany(Alberti 1961); Senonian, Russia (Vozzhennikova 1967); Cenomanian-Turonian, Gabon, Africa (Boltenhagen 1977); Albian-Campanian, Australia(Cookson & Eisenack 1958, 1960, Ingram 1968, Norvick 1876, Morgan 1980); Cenomanian-Campanian, Canada (Manum & Cookson 1964, Davey 1970, Williams 1975, Harker 1979, Singh 1983). Santonian-Campanian, North Dakota(Harker 1979); Cenomanian-Campanian, Wyoming(Griggs 1970, Sulkoske 1975, Harker 1979, Nichols & Jacobson 1982); Cenomanian-Turonian, Arizona(Jones 1976); Cenomanian-Maastrichtian, Texas(Zaitzeff & Cross 1971, Harker 1979); Coniacian, Alabama(Leopold and Pakiser 1964); Campanian, New Jersey(May 1980); Albian-Cenomanian, North Atlantic Ocean(Habib 1970, 1972); Cenomanian-Coniacian, deep sea core west of Angola, Africa(Morgan 1978).

Pervosphaeridium brevispinum(Norvick 1976) Below 1982

1976 ?*Exochosphaeridium brevispinum* Norvick, p.53, pl.4,
fig.5-6.

1982 *Pervosphaeridium brevispinum*(Norvick) Below, p.27.

Previously reported occurrences. Cenomanian.

Pervosphaeridium cenomaniense(Norvick 1976) Below 1982

1976 ?*Exochosphaeridium cenomaniense* Norvick, p.52-53, pl.4, fig.4, 8.

1982 *Pervosphaeridium cenomaniense*(Norvick) Below, p.27.

Previously reported occurrences. Cenomanian.

Pervosphaeridium pseudohystrichodinium(Deflandre 1937)
Yun 1981

1937 *Hystrichosphaeridium pseudohystrichodinium* Deflandre, p.25, pl.12, figs.3-4.

1966 ?*Exochosphaeridium pseudohystrichodinium*(Deflandre 1937) Davey et al., p.166.

1969 *Exochosphaeridium pseudohystrichodinium*(Deflandre 1937) Davey, p.163, pl.11, figs.4-5.

1981 *Pervosphaeridium pseudohystrichodinium*(Deflandre 1937) Yun, p.29.

Previously reported occurrences. late Albian to Eocene. Cenomanian-Campanian, Canada(Harland 1973, Singh 1983); Paleocene, California (Drugg 1967); Cenomanian-Santonian, England(Clarke & Verdier 1967, Davey 1969); Albian-Senonian, France(Deflandre 1937, Davey & Verdier 1973); Cenomanian-Senonian, Italy(Serpagli 1964, Corradini 1973); Campanian-Maastrichtian, Poland(Gorka 1963); Albian, Rumania (Baltes 1967); Danian, Germany(W.Wetzel 1952); Cenomanian-Turonian, Australia (Norvick 1976).

Prolixosphaeridium conulum Davey 1969

Previously reported occurrences. Cenomanian.

Pterodinium cingulatum(O.Wetzel 1933) Below 1981

Previously reported occurrences. Senonian.

Scriniodinium ?compaluna Gocht 1959

Previously reported occurrences. Hauterivian.

Sentusidinium eisenackii(Boltenhagen 1977) Lentin & Williams 1981

1977 *Tenua eisenackii* Boltenhagen, p.56-58, pl.5, fig.5-8.

1981 *Sentusidinium eisenackii*(Boltenhagen) Lentin & Williams,

p.253.

Previously reported occurrences. Cenomanian-Turonian.

Spiniferites ancoriferus Cookson & Eisenack 1974

Previously reported occurrences. middle to late Cretaceous.

Spiniferites cornutus laevimurus (Davey & Williams
1966) Lentin & Williams 1973

1966 *Hystrichosphaera cornuta* var. *laevimura* Davey & Williams,
p.44-45, pl.4, fig.5.

1973 *Spiniferites cornutus laevimurus* (Davey & Williams) Lentin
& Williams, p.127.

Previously reported occurrences. early Eocene.

Spiniferites multibrevis (Davey & Williams 1966) Below
1982

1966 *Hystrichosphaera ramosus* var. *multibrevis* Davey &
Williams, p.35-36, pl.1, fig.4; pl.4, fig.6; test-fig.9.

1982 *Spiniferites multibrevis* (Davey & Williams) Below, p.35.

Previously reported occurrences. Hauterivian.

Spiniferites ramosus ramosus (Ehrenberg 1838) Deflandre
1937

Previously reported occurrences. late Cretaceous.

Spiniferites ramosus reticulatus (Davey & Williams
1966) Lentin & Williams 1973

Previously reported occurrences. Cenomanian.

Subtilisphaera deformance (Davey & Verdier 1973)
Stover & Evitt 1978

1971 *Deflandrea asymmetrica* Davey & Verdier, p.39, pl.2,
figs.4, 6. (invalid as a junior homonym of *Deflandrea*
asymmetrica Wilson 1967, p.62).

1973 *Deflandrea deformance* Davey & Verdier, p.197 (as nom.
nov. pro *Deflandrea asymmetrica* Davey & Verdier 1971,
p.39).

1978 *Subtilisphaera deformance*(Davey & Verdier 1973) Stover & Evitt, 1978, p.239.

Previously reported occurrences. Albian to middle Cenomanian. Albian-Cenomanian, Canada (Bujak & Williams 1978, Singh 1983); Albian, France(Davey & Verdier 1971).

Subtilisphaera foliacea Eisenack & Cookson 1960

Previously reported occurrences. Turonian-middle Senonian.

Surculosphaeridium longifurcatum(Firtion 1952) Davey et al. 1966

Previously reported occurrences. Cenomanian.

Systematophora cretacea Davey 1979

Previously reported occurrences. Albian.

Tanyosphaeridium salpinx Norvick 1976

1976 *Tanyosphaeridium salpinx* Norvick in Norvick & Berger, p.63, pl.9, fig.10.

Previously reported occurrences. Aptian-Cenomanian, Canada (Singh 1971, 1983); Aptian-Cenomanian, Australia(Norvick 1976, Morgan 1980).

Trichodinium castaneum(Deflandre 1935) Clarke & Verdier 1967

1935 *Palaeoperidinium castanea* Deflandre, p.229, pl.6, fig.8.

Previously reported occurrence. Valanginian to Maastrichtian. Valanginian-Campanian, Canada(McIntyre 1974, Williams 1975, Bujak & Williams 1978, McIntyre & Brideaux 1980, Singh 1983); Berremian-Turonian, England(Cookson & Hughes 1964, Clarke & Verdier 1967, Davey 1974, Duxbury 1980); Aptian-?Senonian, France(Deflandre 1936, Davey 1969, Davey & Verdier 1971, 1974); Cenomanian-Maastrichtian, Europe(Foucher 1979); Aptian-Turonian, Australia(Cookson & Eisenack 1962, Norvick 1976, Morgan 1980); Albian, India(Jain 1977).

Xenascus ceratioides(Deflandre 1937) Lentin &
Williams 1973

Previously reported occurrences. Senonian.

Xiphonophorum alatum(Cookson & Eisenack 1962)
Sarjeant 1966

1962 *Hystriochodium alatum* Cookson & Eisenack, p.487, pl.2,
figs.1-4.

1966 *Xiphonophorum alatum* (Cookson & Eisenack 1962) Sarjeant,
p.147-148, pl.16, fig.11.

Previously reported occurrences. Aptian-Cenomanian, Australia
(Cookson & Eisenack 1962); Albian-Cenomanian, England (Cookson
& Hughes 1964, Davey 1970).

APPENDIX B
LITHOLOGIC DESCRIPTIONS OF SAMPLES

(1) Blue Point, Arizona (A1)

Sample 43+: Laminar, olive gray, silty, calcareous shale
Sample 43: Laminar, light olive gray, silty, calcareous shale
Sample 41: Laminar, olive black, slightly silty, calcareous shale
Sample 39: same as above
Sample 36: same as above
Sample 34: same as above
Sample 31: Laminar, olive black, calcareous shale
Sample 28: Olive black, calcareous claystone
Sample 24: Olive black, calcareous claystone
Sample 21: Dark yellow brown, calcareous claystone
Sample 18: Yellow brown, silty, calcareous claystone
Sample 15: Gray, silty, calcareous claystone
Sample 11: Light olive gray, silty, calcareous shale
Sample 8: Light olive gray, silty, calcareous claystone
Sample 5: Light olive gray, silty, calcareous claystone
Sample 3: Light olive gray, silty, slightly calcareous claystone

(2) Pueblo, Colorado (C2)

Sample 46: Olive gray, marly shale
Sample 44: Dark gray, shaly marlstone
Sample 41: Olive gray, marly shale
Sample 37: Olive gray, marly shale
Sample 34: Olive gray, marly shale
Sample 31: same as above
Sample 29: Dark olive gray, marly shale
Sample 27: Olive gray, marly shale
Sample 23: Light olive gray, marly shale
Sample 18: Medium light gray, shaly marlstone
Sample 16: Medium dark gray, shaly marlstone
Sample 14: Light olive gray, slightly silty, marly shale
Sample 11: Light olive gray, marly shale
Sample 8: Light olive gray, calcareous claystone

(3) Boulder, Colorado (C4)

Sample 2030: Dark olive gray, calcareous shale
 Sample 1946: Olive gray, calcareous shale
 Sample 1852: Laminar, dark olive gray, calcareous shale
 Sample 1777: Laminar, olive gray, calcareous shale
 Sample 1697: Olive gray, calcareous shale
 Sample 1640: Dark olive gray, calcareous shale
 Sample 1580: Laminar, olive black, calcareous shale
 Sample 1510: Laminar, olive black, calcareous shale
 Sample 1463: Olive black, calcareous shale
 Sample 1372: Laminar, olive black, calcareous shale
 Sample 1302: Olive black, calcareous shale
 Sample 1242: Laminar, dark olive gray, calcareous shale
 Sample 1149: Laminar, olive black, calcareous shale
 Sample 1094: Laminar, olive gray, calcareous shale
 Sample 974: Laminar, olive gray, calcareous shale
 Sample 894: Laminar, olive black, calcareous shale
 Sample 740: Laminar, olive black, calcareous shale
 Sample 635: Laminar, olive black, calcareous shale
 Sample 520: Laminar, olive black, calcareous shale
 Sample 395: Laminar, olive black, calcareous shale
 Sample 275: Laminar, olive black, calcareous shale
 Sample 200: Dark olive gray, calcareous shale
 Sample 100: Laminar, olive black, calcareous shale
 Sample 42: Laminar, olive black, calcareous shale
 Sample 36: olive black, calcareous shale
 Sample 34: Olive black, calcareous shale
 Sample 33: Laminar, olive black, calcareous shale
 Sample 28: Laminar, olive black, calcareous shale
 Sample 26u: Brown black, calcareous marlstone
 Sample 26m: same as above
 Sample 261: same as above
 Sample 25: Brown black, calcareous shale
 Sample 23: Dark gray, calcareous shale
 Sample 21: Laminar, dark brown, calcareous shale
 Sample 19: Laminar, dark brown, calcareous shale
 Sample 15: Laminar, dark brown, calcareous shale
 Sample 13: Laminar, dark brown, calcareous shale
 Sample 11: Laminar, dark brown, calcareous shale
 Sample 6: Laminar, olive black, calcareous shale
 Sample 3: Gray green, calcareous shale
 Sample -034: Laminar, olive black, calcareous shale
 Sample -107: Laminar, dark olive gray, calcareous shale
 Sample -204: Laminar, olive black, calcareous shale
 Sample -297: Laminar, dark olive gray, calcareous shale
 Sample -391: Laminar, olive black, calcareous shale
 Sample -525: Laminar, olive black, calcareous shale
 Sample -690: Laminar, dark olive gray, calcareous shale

** the number for those sample above sample 42 represents
 the thickness (cm) above the bed 42; and for those below

sample 3 represents the thickness (cm) below the bed 1.

(4) Ponca, Nebraska (N1)

Sample 36: Medium olive gray, chalky limestone
 Sample 33: same as above
 Sample 28: Olive gray, shaly, chalky limestone
 Sample 24: same as above
 Sample 19: Olive gray, calcareous claystone
 Sample 16: Olive gray, chalky shale
 Sample 14: Olive gray, chalky shale
 Sample 12: Olive gray, chalky shale
 Sample 8: Olive gray, chalky shale
 Sample 6: Olive gray, chalky shale

(5) Carthage, New Mexico (NM2)

Sample 42: Laminar, dark olive gray, calcareous shale
 Sample 38: Laminar, dark olive gray, calcareous shale
 Sample 35: Laminar, dark olive gray, calcareous shale
 Sample 33: Laminar, dark olive gray, marly shale
 Sample 30: Laminar, dark olive gray, calcareous to marly shale
 Sample 28: Laminar, olive black, slightly silty, marly shale
 Sample 27: Laminar, olive black, slightly silty, marly shale
 Sample 25: Laminar, dark olive gray, marly shale
 Sample 23: Medium olive gray, shaly limestone
 Sample 21: Medium olive gray, slightly silty, calcareous shale
 Sample 18: Laminar, olive black, calcareous shale
 Sample 16: Laminar, dark olive gray, calcareous shale
 Sample 13: Laminar, olive gray, calcareous shale
 Sample 10: Laminar, olive gray, calcareous shale
 Sample 8: Laminar, olive gray, slightly silty, calcareous shale
 Sample 6: Laminar, olive gray, calcareous shale
 Sample 4: Laminar, olive gray, slightly silty, calcareous shale
 Sample 1: Dark olive gray, calcareous claystone

(6) Dallas, Texas (T1)

Sample 35: Laminar, olive gray, calcareous shale
 Sample 34: same as above
 Sample 32: Laminar, dark olive gray, calcareous shale

Sample 31: Laminar, dark green gray, shale
 Sample 29: same as above
 Sample 27: same as above
 Sample 25: same as above
 Sample 23: same as above
 Sample 21: same as above
 Sample 18: Laminar, green black, shale
 Sample 15: Laminar, dark olive gray, shale
 Sample 12: Laminar, dark green gray, slightly calcareous shale
 Sample 10: same as above
 Sample 6: Laminar, dark green gray, slightly calcareous shale
 Sample 4: Laminar, dark green gray, slightly calcareous shale
 Sample 1: Laminar, dark green gray, slightly calcareous shale

(7) Wahweap Wash, Utah (U1)

Sample 50: Dark olive gray, calcareous shale
 Sample 46: Dark gray, calcareous shale
 Sample 40: Laminar, olive black, calcareous to marly shale
 Sample 35: same as above
 Sample 31: same as above
 Sample 27: same as above
 Sample 25: Laminar, olive black, calcareous shale
 Sample 23: Medium gray, calcareous marlstone
 Sample 20: Olive black, slightly silty, calcareous claystone
 Sample 18: same as above
 Sample 16: Light olive gray, slightly silty, calcareous claystone
 Sample 14: Dark olive gray, calcareous claystone
 Sample 12: same as above
 Sample 8: Laminar, olive gray, slightly silty, calcareous shale
 Sample 5: Dark olive gray, silty, calcareous claystone

(8) Egg Nog, Utah (U2)

Sample 31: Laminar, medium olive gray, calcareous shale
 Sample 27: Light olive gray, silty, calcareous claystone
 Sample 24: Olive gray, slightly silty, calcareous claystone
 Sample 21: Olive gray, slightly silty, calcareous claystone
 Sample 18: Laminar, olive gray, slightly silty, calcareous shale
 Sample 16: Medium lighy gray, calcareous shale to claystone
 Sample 14: Medium gray, calcareous claystone

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